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UNIVERSITY OF ALBERTA

SECONDARY SEDIMENTARY STRUCTURES IN THE CYPRESS HILLS AND WOOD MOUNTAIN FORMATIONS, SOUTHWESTERN SASKATCHEWAN

by

MICHELLE M. PERRAS

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE OF MASTER OF SCIENCE

DEPARTMENT OF GEOLOGY

EDMONTON, ALBERTA

FALL, 1991

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FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled SECONDARY SEDIMENTARY STRUCTURES IN THE CYPRESS HILLS AND WOOD MOUNTAIN FORMATIONS, SOUTHWESTERN SASKTCHEWAN submitted by Michelle M. Woytenko in partial fulfilment of the requirements for the degree of MASTER of SCIENCE.

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Date Oct 8th 1991

Abstract

Tertiary gravels form the caprock of the Cypress Hills and Wood Mountain uplands in southwestern Saskatchewan. They are left as remnants of a semi-arid alluvial plain. Although unglaciated during the Late Wisconsin ice advance, these gravels were altered by processes during this event.

Eleven facies and three secondary sedimentary structures were identified. The facies distribution is controlled by the depositional environment and ranges from coarse gravelly facies in the western portion of the study area (proximal to the source) to dominantly sandy facies to the east. Typical section profiles include interbedded clast-supported facies fining upwards into finer, matrix-supported gravel facies. Contorted units, pillars, and reoriented units were identified as secondary structures. These structures are formed from pressure variations and confinement as the overlying sediment became frozen under permafrost conditions during glaciation.

The distribution of the secondary structures is controlled by the distribution of the primary lithofacies and the Late Wisconsin ice limit. Structures are more common in the western portion of the study area where coarser facies are more common. All observed structures are located beyond the mapped limits of the Late Wisconsin ice.

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Table of contents

Chapter

1. Introduction.....1
   1.1. Objectives.....1
   1.2. Climate.....3
   1.3. Bedrock Geology.....4
   1.4. Quaternary Geology.....4
   1.5.Previous Work.....5

2. Field Methodology and Laboratory Techniques.....8

3. Interpretation of Sedimentary Textures and Structures.....10
   3.1. Alluvial Deposits.....10
      3.1.1. Size Distribution and Sorting.....10
      3.1.2. Grading.....11
      3.1.3. Stratification.....11
      3.1.4. Fabric.....13
      3.1.5. Clast orientation.....13
      3.1.6. Clast Shape and Orientation.....14
   3.2. Flows.....15
      3.2.1. Debris flows.....16
      3.2.2. Debris flow textures.....17
      3.2.3. Grain Flows.....18
      3.2.4. Slumps and Slides.....18
      3.2.5. Fabric of Flows.....19

4. Influence of Frost Action.....21
   4.1. Frost Push.....21
   4.2. Frost Pull.....22
   4.3. Inclined Freezing Front.....23
   4.4. Reorientation Process.....26
   4.5. Factors affecting the growth of ice lenses.....29
      4.5.1. Soil Texture.....29
      4.5.2 Pore size.....31

4.5.3 Mineralogy.....32
4.5.5 Clast Shape.....33

5. Descriptions and Interpretations.....34
5.1. Lithofacies Descriptions and Interpretations 34
Lithofacies 1
Horizontally stratified, pebble-matrix supported gravel.....36
Lithofacies 2
Structureless, matrix supported gravel 38
Lithofacies 3
Horizontally stratified, matrix supported gravel.....40
Lithofacies 4
Structureless, clast supported gravel 41
Lithofacies 5
Crudely stratified, clast supported gravel 41
Lithofacies 6
Trough cross-stratified, clast supported gravel 42
Lithofacies 7
Openwork gravel.....43
Lithofacies 8
Massive to horizontally laminated sands 45
Lithofacies 9
Type B Climbing Rippled Sands 45
Lithofacies 10
Trough Cross-bedded Sands 46
Lithofacies 11
Massive silts with rare pebbles 47

6. Description and Interpretation of secondary structures.....48
6.1. Contortions 48
6.1.1. Descriptions.....48
6.1.2. Interpretation of Contortions.....56
6.2. Pillars 61

6.2.1. Descriptions.....61
6.2.2. Interpretation.....66
6.3. Reoriented Units 74
6.3.1.Descriptions 75
6.3.1.1 Matrix 75
6.3.1.2 Clasts 81
6.3.2.Interpretation 81

7. Distribution of Secondary Sedimentary Structures.....89
7.1. Distribution of Contorted Units 89
7.2. Distribution of Pillars 90
7.3. Distribution of Reoriented Units 91

8. Conclusions.....93

References.....97

List of Figures

Figure 1.1

Location of the study area..... 2

Figure 1.2

Distribution of secondary structures in the study area..... 6

Figure 5.1

Locations of sections in the study area..... 35

Figure 5.2

Average clast dip versus silt content in reoriented units..... 78

Figure 5.3

Average clast dip versus silt and clay content in reoriented units..... 79

Figure 5.4

Textural comparison of similar depth samples with varying clast dips..... 80

Figure 5.5

Clast shape versus orientation..... 83

List of Plates

Plate 5.1
Typical exposure of lithofacies 1..... 37
Plate 5.2
Typical exposure of lithofacies 7..... 44
Plate 5.3
Contortions..... 49
Plate 5.4
Panoramic of Resch section showing contortions..... 50
Plate 5.5
Close up of the contortions at Resch section..... 51
Plate 5.6
Contortions at Center Block section..... 53
Plate 5.7
Contortions in sand facies at Ravenscrag section... 55
Plate 5.8
Contortions at Police Point section..... 57
Plate 5.9
Pillar exposed at Ravenscrag section..... 63
Plate 5.10
Sands at the base of the Ravenscrag pillar..... 64
Plate 5.11
Pillar exposed at Rockglen section..... 65
Plate 5.12
Typical exposure of reoriented unit..... 77

Chapter 1

Introduction

Investigations of the Cypress Hills and Wood Mountain uplands began with Dawson (1875). He recognized that the highest surfaces of these plateaus were not covered by glacigenic sediment. The area has continued to interest workers as it is the only place in the Canadian Prairies that remained unglaciated. This study provides information on the effects of glaciation on the unglaciated areas. Emphasis is placed on the description and interpretation of secondary sedimentary structures.

The Cypress Hills region encompasses the southeastern corner of Alberta and the southwestern corner of Saskatchewan (108° to 110° W. Longitude) (Figure 1.1). The Cypress Hills are a series of southeastward sloping, flat topped, gravel covered plateaus which reach a maximum elevation of 1450 m near Elkwater, Alberta. These plateaus, locally known as the west, center and east blocks, are truncated abruptly by an escarpment along its northern flank, and stands 150 m above the gently northeastward sloping plain. To the south, the plateaus are bordered by a series of narrow, ill defined, and dissected steppes. North of the plateau, the regional physiography is characterized by gently undulating plains with hummocky terrain. Similar topography is found in the Wood Mountain area (104° to 108° W. Longitude) (Figure 1.1), situated east of the Cypress Hills.

1.1.Objectives

The objectives of this study are:

1) to determine the distribution of the secondary sedimentary structures in the study area;

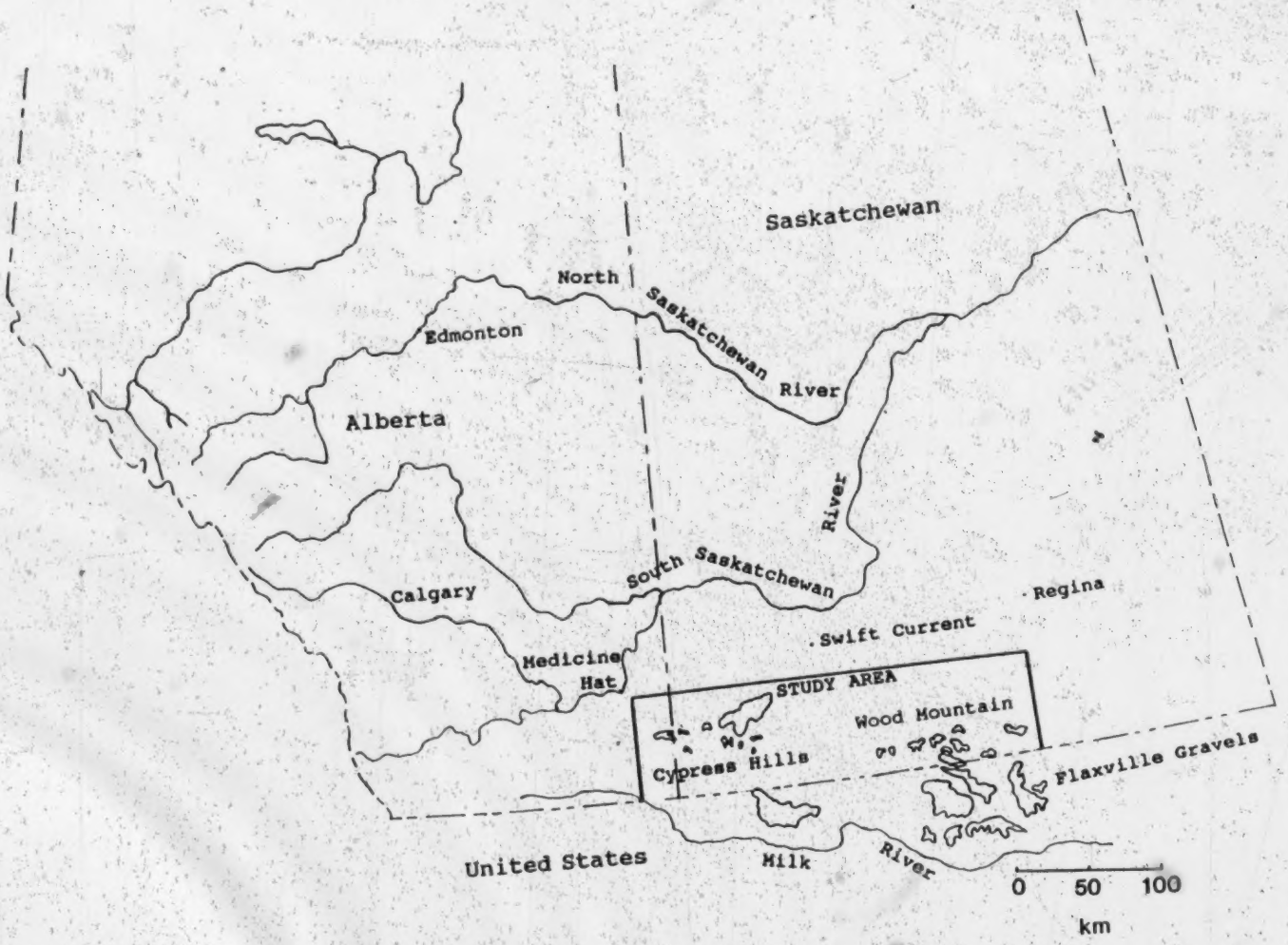


FIGURE 1.1 LOCATION OF STUDY AREA

2) to determine the processes responsible for the formation of the secondary sedimentary structures observed in the gravels and sands;

3) to determine a chronology of events leading to the development of the secondary structures;

and 4) to determine if primary characteristics influence the formation of secondary structures.

1.2.Climate

Southeastern Alberta and southwestern Saskatchewan have a semi-arid climate with hot, relatively dry summers and cold winters. The lowland areas are classified as middle latitude steppes. Although a considerable distance east of the Rocky Mountains, the area is influenced by warm chinook winds at irregular intervals throughout the year. Southwesterly winds prevail during the winter months but are less frequent in the summer (Kendrew and Currie, 1955).

The lowland areas are typified by mean annual temperature varying from 4.1° to 5.5°C (Longley, 1972). The mean annual precipitation is 350mm, a value considerably less than the average annual potential evapotranspiration loss of 571mm (Borneuf, 1976).

On the upland areas, a cool boreal climate prevails especially on the north and northwest facing slopes. Annual precipitation exceeds 400mm for the Cypress Hills Plateaus (Longley, 1972) varying in recent years from 327.6mm in 1988 to 718.6mm in 1986 (Appendix 1). The upland areas have a mean annual temperature of 2.5°C with a range of temperature fluctuation from -41°C to 30°C.

Precipitation and temperature data are given in Appendix 1. Generally, the eastern half of the area receives less precipitation and records higher temperatures than the western half. In the extreme southwest the annual precipitation may not exceed 240mm. The upland areas receive more precipitation than the surrounding prairies.

1.3. Bedrock Geology

The weakly-consolidated, sedimentary Paleocene Ravenscrag Formation and Upper Cretaceous Frenchman Formation are considered bedrock and underlay the Cypress Hills and Wood Mountain Formations throughout the study area. These Formations crop out only on the flanks of the plateaus and the lowlands between the plateaus are underlain by older sediments. A disconformity separates the Frenchman Formation from the underlying, more extensive Battle Formation. The Ravenscrag and Frenchman Formations dip gently to the northeast.

The Ravenscrag Formation is interpreted as a fluvial deposit (Williams and Dyer, 1930), and consists of sandstone with interbedded silts. Rahmani and Lerbekmo (1975) interpreted the Frenchman Formation as a fluvial deposit and it consists of sandstone with rare coal and bentonite beds.

1.4. Quaternary Geology

The Quaternary geology of the Cypress Hills region was first investigated by Dawson (1875, 1885) and McConnell (1885). These researchers identified the major physiographic features of the region, and recognized the presence of glacigenic sediments on the lowland and their absence on the Cypress Hills plateaus. Christiansen (1959, 1965, 1979) recognized three glacial deposits which have

been assigned the ages of Pre-Sangamon, Early Wisconsin, and Late Wisconsin. Klassen and Vreeken (1985, 1987; Klassen, 1984), identified two tills which were interpreted as being the result of Late Wisconsin ice lobes although a pre-Late Wisconsin glaciation in the area is believed to have occurred based on the degree of erosion, weathering of glacial erratics, and soil development. Most recently, the study area has been interpreted as being affected by a single, Wisconsin glaciation (John J. Kulig, Department of Geology, University of Alberta, personal communication, 1991). Although the controversy between the number of glaciations, the author believes that there was only a single, Late Wisconsin glaciation in the study area. The limit of glaciation is shown in Figure 1.2.

1.5.Previous Work

The Cypress Hills Formation was first investigated by McConnell (1885). He believed that the surface of the plateaus were covered by a Miocene fluvial conglomerate with a source in the Rocky Mountains. Vonhof (1969) studied the sedimentology of the Cypress Hills and Wood Mountain Formations. He interpreted the deposits as a meandering stream environment. Leckie and Cheel (1989) interpreted the Cypress Hills Formation as a semi-arid braid plain depositional environment. The Cypress Hills Formation is interpreted to represent continuous deposition from Late Eocene to Early Oligocene, spanning the period 44-35 Ma (Leckie and Cheel, 1989). Vonhof (1969) interpreted the age of the Wood Mountain Formation as Middle to Late Miocene. Although the Wood Mountain Formation is younger, it may represent a downstream facies change from the coarser grained Cypress Hills Formation.

Many authors have indicated that the top few meters of the Cypress Hills formation has been reworked (Westgate 1965,1968,1972; Jungerius 1966,1967,1969; Jungerius and

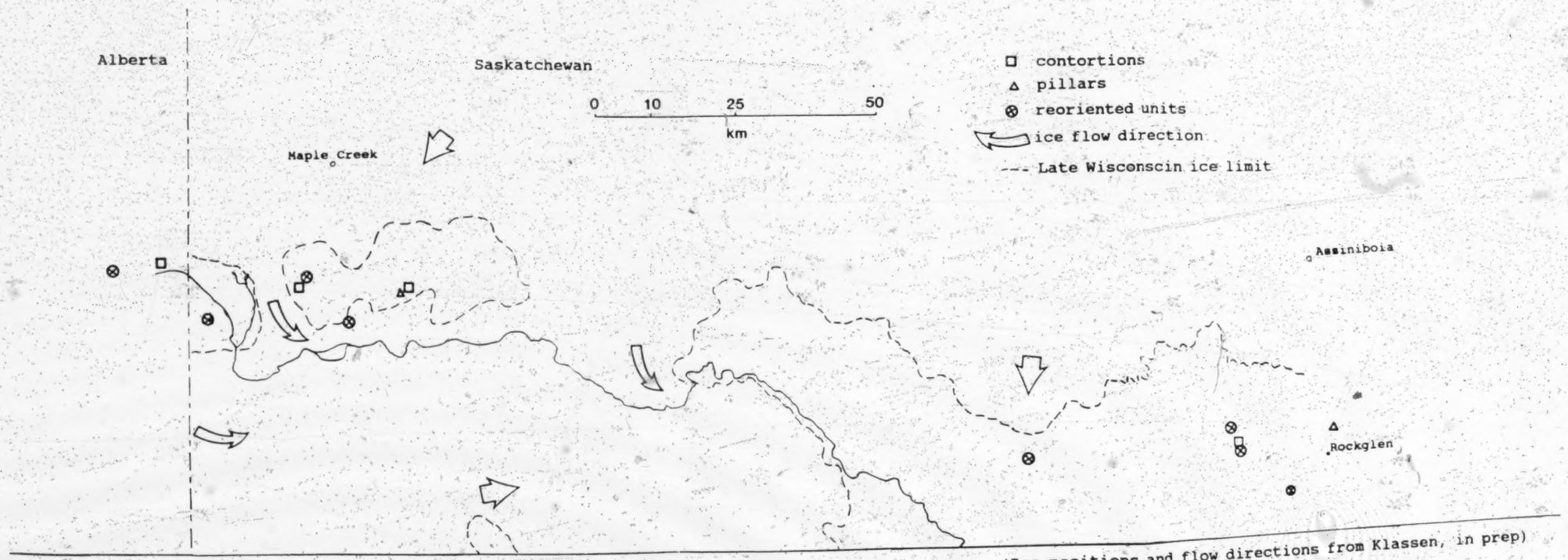


FIGURE 1.2 Late Wisconscin Ice Limits and Distribution of Secondary Sedimentary Structures

(Ice positions and flow directions from Klassen, in prep)

Murcher 1972; Catto 1981). Possible processes speculated include reworking by frost, overburden pressure, and colluvial or glacial deposition. Jungerius and Murcher (1972) believed that frost was not necessarily the exclusive agent. They concluded that the disturbance may be the result of direct pressure from glaciers. Westgate (1968) saw this unit as poorly sorted and in places coarse grained, characteristics atypical of loess. He explained this feature by frost mixing some of the conglomerate up into the loess. Catto (1981) also disagreed with Jungerius and Murcher (1972) concluding that the Cypress Hills was unglaciated and the till-like material near the crest of the hills was colluvial in origin.

Chapter 2

Field Methodology and Laboratory Techniques

Outcrops of the Cypress Hills and Wood Mountain Formations in southwestern Saskatchewan and southeastern Alberta were measured and described on a bed by bed basis (Figure 1.1). A facies scheme was developed and modified during the field season as new sediment types were observed. Sediment sampling in individual beds included a 1 kg sample for grain size analysis, measurements of the a-, b-, and c-axes of at least 200 clasts, and fabric measurements on the a-b axial plane of elongate clasts. Samples were taken for grain size analysis and clast lithology.

Measured beds have been divided into sedimentary facies, with the aid of the Udden-Wentworth grain size scale and stratification type. This is similar to the approach used by Davies and Walker (1974). Beds showing internal stratification were placed in facies based on stratification type. Basal contacts were described as flat, undulose, or scoured and are related to the depth of penetration of the overlying bed into the underlying bed(s). Flat contacts show no evidence of erosion, forming a smooth contact with the underlying bed. Undulose contacts are characteristically wavy with scour depths less than 5cm. Scoured contacts are irregular surfaces with scour depth less than 50cm. Loaded bedding contacts are also included. Loaded contacts are not related to an erosive event but simply shallow (5-8 cm) to large (25-65 cm) deformational features. The term "structureless" herein follows the useage of Hein and Walker (1982) and applies to those beds showing some type of grading, but are devoid of traction sedimentary structures. The term "massive" applies to beds showing no grading or tractional features.

Textural samples were analyzed using the hydrometer method specified in ASTM D422 (1963) and sieved following the procedures in ASTM D423 (1963). The particle size distributions were plotted and grain size diameters were calculated using formulas and procedures given in Folk and Ward (1957) and Folk (1966) with the aid of the computer program JAY.HYD developed by Gerry Cyre of the Civil Engineering Department at the University of Alberta and modified by Michael Woytenko.

Fabrics were determined using sample sizes of a minimum of 25 clasts. The clasts were >2.5 cm (a-axis) and had a:b axial ratios of at least 2:1. The stereonets were plotted using STEREO, a computer program developed by D.B. McEachran. The eigen vectors are an indication of the dominant orientation of the clasts while the eigen values give an indication of the strength of the eigen vectors.

Chapter 3

Interpretation of Sedimentary Textures and Structures

The purpose of this chapter is to outline the current interpretations of gravel sedimentary textures and structures. This will aid in the interpretation of the lithofacies in Chapter 5. The topic will be limited in scope to descriptions of those textures and structures found in alluvium of braided streams and alluvial fans, the current interpretation of the depositional environment of the Cypress Hills and Wood Mountain Formations.

Texture, structure and pebble fabric analysis of alluvial fans and braided streams and, to a lesser extent, subaerial debris flows are documented in studies of alluvial fans (Bull, 1970; Fisher, 1971; Enos, 1977), modern braided rivers (Smith, 1974; Hein and Walker, 1977; Church and Jones, 1982; Hein, 1984; Southard *et al*, 1984) and sandurs (Rust, 1972; Bradley *et al*, 1972; Fahnestock and Bradley, 1973; Boothroyd and Ashley, 1975).

3.1. Alluvial Deposits

3.1.1. Size Distribution and Sorting

Fan-head and fan-channel gravels are angular, poorly-sorted gravels which are texturally similar to the weathered bedrock from which they are derived. The degree of textural sorting of both the clasts and matrix is related to local provenance and minimal reworking of the gravel. Well documented down-fan trends include a decrease in the mean and maximum grain sizes (Blissenbach, 1954). The most accurate measure of progressive clast-size reduction down-fan is the measure of the maximum intermediate (b) axis which shows an exponentially decreasing relationship with transport distance. Grain-size changes of gravel alluvium in major valleys and on alluvial plains occur as different

size clasts respond to shear stress on the bed by different modes and rates of movement. For gravel clasts the most common type of movement is rolling of the clast about a pivot formed by other clasts (Middleton and Southard, 1984). Sorting by this tractional movement is less efficient than sorting by suspension, but in alluvium, varying hydraulic conditions act to sort all of the sediment load into different parts of the system. In gravel-dominated braided rivers, downslope decrease in grain size occurs over a considerable distance, generally an order of magnitude greater than that required on alluvial fans.

3.1.2.Grading

Grading refers to progressive clast-size change within an individual bed. Both normal and inverse grading in fluvial gravels suggests that processes of transport and deposition sort bedload into different sizes. Normal grading of clast sizes indicates that the flow strength was diminishing during deposition. Grading in gravels can be used to infer changes in either forces acting on the bed or variations in flow velocity. However, major problems exist in the selection of criteria by which grain size can be used to infer paleoflow parameters. Inverse grading can occur by infiltration or (on alluvial fans) by sieving.

3.1.3. Stratification

Sedimentary structures in gravel alluvium are composed of differing layers of clast-supported, matrix-filled gravels of different grain sizes. Less common are those structures characterized by changes in clast-fabric orientation or alternations of sandy matrix-filled gravels with open-work gravel. Harms et al (1982) have shown that in some cases alternating sand and gravel layers may imply only a slight fluctuation in flow strength. Bluck (1976) suggested that pockets of openwork texture may be due to post-depositional

winnowing of the matrix. Smith (1974) interpreted the occurrence of openwork fine granule-size gravel and sandy interbeds with coarse matrix-filled gravels as recording fluctuating discharge levels.

Four major types of stratification occur in coarse-grained braided alluvium: 1) horizontal stratification; 2) planar tabular cross-bedding; 3) trough cross-bedding; and 4) large scale cross-bedding and lateral accretion cross-bedding. Horizontal stratification is most often described in outwash-plain settings (Hein and Walker, 1977) dominated by rapidly shifting braid channels with longitudinal or diagonal braid bars that develop into thin diffuse gravel sheets. Similar crude stratification occurs in alluvial fans and is attributed to deposition from unchannelized sheet floods (Bull, 1970). Growth of transverse bars and their associated slip-faces in distal reaches of rivers result in the formation of planar-tabular or wedge-shaped cross-beds on the downstream margin of bars (Smith, 1970). Trough cross-stratification is the product of migrating dunes in channels (Harms et al, 1982).

Attempts have been made by Rust (1977), Hein (1984), and Smith (1985) to clarify the relationship between the origin and types of stratification in modern rivers. Development of crudely-defined horizontal to low-angle stratification is related to the progradation of longitudinal and diagonal bar margins where high angle planar tabular cross strata are associated with the formation of steep bar margins. The role of rapid decreases in water and sediment discharge in the development of cross-stratification was discussed by Hein and Walker (1977). Their model suggests that angle of repose cross-stratification arises when fluid discharge and sediment discharge across an incipient bar core both decline rapidly after formation of a diffuse gravel sheet. Under these reduced fluid and sediment discharges, vertical

aggradation dominates over downstream migration, resulting in the development of angle of repose foresets. By contrast, low angle stratification forms when the rate of migration is about the same as the rate of vertical aggradation, as in the case of longitudinal or diagonal bars. In these cases, no angle of repose foresets are developed, and the bars form riffle margins. Upon progradation, the riffle margins produce horizontal to low angle stratification (Hein and Walker, 1977; Hein, 1984).

3.1.4. Fabric

The term fabric refers to the arrangement in a rock of the various components, including the packing and orientation of clasts. Clast supported gravels and conglomerates composed of pebbles with long axes greater than 2 cm display two main types of clast orientation: 1) a-axis transverse to flow, b-axis imbricate upstream; and 2) a-axis parallel to flow, a-axis imbricate upstream (Harms et al, 1982). Johansson (1965) experimentally showed that the a-axis transverse to flow and b-axis imbricate pattern is the result of clasts rolling along the bed. This has also been observed in modern river deposits (Rust, 1972). The less common, a-axis parallel to flow, a-axis imbricate has only been documented in "deep water turbidite-like" conglomerates (Harms et al, 1982). This fabric pattern does not develop as a result of the clasts rolling along the bed but may indicate transport processes that maintain pebbles and cobbles in dispersion above the bed. These processes may also occur in some alluvial fan floods (Harms et al, 1982).

3.1.5. Clast orientation

Pebble orientation is an important paleocurrent indicator for coarse fluvial gravels and conglomerates and is commonly measured as the trend and plunge of the a-b plane of disc-

shaped pebbles, and the azimuth and dip of the a-axis of elongate pebbles. In high velocity (high energy) river deposits, the most frequent orientation is transverse to flow with an upstream imbrication of about 10° to 30°. Lower angles of imbrication are expected on beaches (Blatt et al, 1980). The range of dip angles for fluvial deposits is 6° to 30°, with modal values of approximately 20°. Graded turbidite deposits (Kopstein, 1954) were found to have an average dip of 25° and beach deposits exhibited dips up to 45° (Krumbein, 1939).

3.1.6. Clast Shape and Orientation

Based on casual observation, Becker (1893) noted that flat pebbles shingle upstream and elongate pebbles were found lying across the channel. Twenhofel (1932) found that pebbles were placed on a sand bed in a current too weak to transport them, the washing of the sand beneath the clasts was sufficient to rotate the clasts until they attained a position with their longest axis perpendicular to current. Cailleux (1945) found that upstream inclination of pebbles tends to decrease with increasing flatness.

Schlee (1957) conducted flume studies using rod and disc shaped clasts. He found that the disc shaped pebbles formed a more distinctive fabric pattern whereas rods formed a more scattered fabric. The elongate pebble can be considered the limiting case for the discoidal pebble with a much shortened b-axis. When such a clast is normal to current flow, the pebble's a-b plane may act as an obstruction to flow. This leads to washing out of the upstream side of the pebble, and results in a steep downstream dip of the short axis and the long axis remaining horizontal. Johansson (1965) found that the dip of flat pebbles and cobbles in river beds steepens with increased flow velocity and depth, increased clast size, decreased prolateness, and increased rounding. The roundness and shape of the clast influences its resistance

to movement (potential energy). Rust (1972) showed that flat pebbles generate relatively small roller vortices, low lateral velocities, and small scouring cavities because of their low potential energy. Thin pebbles do not tilt upstream as readily as thicker pebbles because they have a relatively larger area of contact with the bed and relatively higher friction. Thick pebbles cause strong cavity scouring and tilt more readily than flat pebbles. Flat particles assume a more stable imbricated position than less discose ones under moderate flow velocity and turbulence. Increasing velocity and turbulence cause diminishing stability, especially for flat particles. The lift forces generated under high turbulence have the greatest influence on flat particles, which have a relatively large surface area. High velocity, turbulent flow also causes decreased regularity of orientation. Flat and elongate particles usually attain a longitudinal position as a result of movement above the bed due to shear generated by flow, or rotating around previously deposited clasts.

3.2. Flows

Middleton and Hampton (1973) classified sediment gravity flows into four groups: debris flows, grain flows, turbidity currents and fluidized flows based on the type of sediment support mechanism during transport. Subaerial debris flows were defined as gravity propelled movements of sediment-water mixtures in which grains are supported above the bed by the strength and buoyancy of a clay/water matrix (Middleton and Hampton, 1973). Lowe (1979, 1982) proposed that sediment gravity flows be classified by their rheological behavior into debris flows with plastic behavior and fluidal flows with viscous fluid behavior (grain flows).

3.2.1. Debris flows

Debris flow refers to a sluggish movement of granular solids, clay and water in response to the gravitational pull (Middleton and Hampton, 1973, 1976). Granular solids are suspended within a clay/water matrix with a density of 2.0-2.5 g/cm3 (Curray, 1966; Middleton and Hampton, 1973). Hampton (1975, 1979) experimentally showed that with increasing grain size the concentration of the clay-water mixture increases. Fine grained sands suspended in debris flows with bulk clay content as low as 2 wt%. Coarse grained sand sediment can move as debris flows if they contain 19 wt% clay or less. The competence of the matrix is a measure of the largest supported grain size, and is controlled by the strength and density of the clay-water mixture (dispersive pressure playing a less important role). Movement of this type of sediment-gravity flow is aided by the low hydraulic permeability of the clayey matrix. This effectively reduces the settling velocity of the granular sediment, enabling the development of sufficient momentum for continued downslope movement (Morgenstern, 1967). Debris flows are capable of sustained movement even on shallow slopes (Curray, 1966).

Deposition from debris flows occurs by mass emplacement when the driving force, gravity, is exceeded by the yield strength of the debris flow. Deposits are characteristically poorly sorted, with a disorganized fabric (Crowell, 1957; Fisher, 1971), and commonly contain out-sized clasts randomly distributed throughout a fine-grained matrix. This concept of debris flow deposits has since been modified to include internally graded beds (both normal and inverse types) (Hampton, 1972, 1975, 1978).

3.2.2. Debris flow textures

Debris flow movement has been observed and modelled by Johnson (1970). During movement of debris flows, a rigid plug riding passively over a basal region experiences applied shear stresses which exceed the fluid yield strength. Sedimentary textures, particularly the grading and sedimentary fabric, depend on whether the flow is turbulent or laminar internally. In plastic debris flows (characterized by high-strengths and viscous interactions between the clasts) inverse grading occurs within the basal shear zone which occurs most commonly as a mud-rich, relatively clast-free region only a few centimeters thick (Shultz, 1984). Grading is rarely documented in ancient deposits, but partially inversely graded beds have been reported (Nemec and Steel, 1984; Gloppen and Steel, 1981). Products of more dilute and "pseudoplastic" fine-grain debris flows would yield inverse-to-normally graded, or ungraded to normally graded deposits in the coarsest flows (Shultz, 1984).

Several different theories attempt to explain the pressures and forces in the basal shear zone that give rise to inverse grading. Bagnold (1954) suggested that the shear zone develops dispersive pressures and has the greatest grain mobility which favors the upward movement of large grains. Fisher (1971) suggested that this zone is a region of upward decrease in pressure gradients (as a result of velocity increases) that displace clasts upward. Middleton (1970) proposed a "kinetic sieve mechanism" during movement of the bed to explain inverse grading. The mechanism depends upon the tendency of smaller particles to fall into the spaces between the larger particles. Thus, the kinetic interactions between the particles during movement displaces the larger particles toward the upper surface of the flow.

3.2.3. Grain Flows

The concept of grain flow arose from the theoretical and experimental work by R.A. Bagnold (1954, 1956), who experimentally measured a dispersive force generated after a concentration of cohesionless grains was sheared. This dispersive pressure results in the upward displacement of the grains against the force of gravity, resulting in a reduction in the yield strength of the granular mass. Downslope movement will begin and continue until the gravitational force is overcome by the yield strength of the mass. Flow of this type requires very high angles of slope in order to sustain the dispersion necessary for movement and grain to grain interactions. Bagnold (1973) and Francis (1973) reported that the inertial-dominated grain flow requires slopes of 32°, however those dominated by the viscous regime require a slope of 37° (Bagnold, 1973; Middleton and Hampton, 1973, 1976).

Grain flow deposits are characteristically sharply bounded, ungraded sandstones (Stauffer, 1967). Basal contacts occasionally show unusual sole markings. When present, lithological rip-up clasts are randomly dispersed.

3.2.4.Slumps and Slides

Slumps and slides refers to mass movements in which the elastic limit of the sediment has not been exceeded (Dott, 1963; Nardin et al, 1979). Movement occurs along discrete internal shear planes when the strength force (that which resists deformation) of an inclined sediment pile is overcome by the gravitational forces. Submarine slumps and slides are capable of movements on slopes with gradient of only 0.5° (Grant-Mackie and Lowry, 1964). Slide deposits show only minor internal deformation, and are typically confined to a narrow region at the base of the bed.

Original sedimentary structures are typically well preserved. Slumps on the other hand show greater internal deformation. Extreme deformation is indicated by the occurrence of elongated pieces of remnant bedding "floating" in a mud or sand matrix. Deposits typically exhibit slump folds whose axial planes are oriented downslope (Helwig, 1970). Basal contacts of both slumps and slides represent a decollement zone and lack the erosional surfaces typical of other sediment gravity flow deposits.

3.2.5. Fabric of Flows

A very poorly organized fabric is characteristic of debris flow deposits. Dispersive pressure in the flow causes clast segregation with the largest clasts moving to the zone of least shear at the top of the flow. The deposit produced should exhibit inverse grading and a sub-horizontal a-b axial clast fabric (Fisher, 1972; Enos, 1977). Viscous flows have lower water contents, move slowly and have irregular clast distributions with many of the clasts possessing vertically-oriented a-axes (Bull, 1977; Pierson, 1980). Fluid laminarly flowing debris flows possess sub-horizontal to horizontal clast imbrications and well-developed clast fabrics (Bull, 1977; Fisher, 1971; Enos, 1977). Lindsay (1978) developed a computer simulation of clast alignment in which shearing in a laminar viscous fluid-flow, aligned the a-axis of the clasts parallel to flow and dipping upstream. This fabric tended to degrade and reappear in a cyclical pattern. Other types of clast orientation have also been mentioned including a-axis or b-axis upflow imbrication (Nemec and Steel, 1984). The number of clast interactions and the relative dispersive pressure strength influences the type of clast imbrication. As these factors differ between flows, clast imbrication patterns can vary considerably (Nemec and Steel, 1984).

Despite observing numerous near vertically-oriented clasts, Kulig (1985) found that the average plunge for debris flow deposits ranged from 0.3° to 25°. Most of these deposits had a multi-modal distribution and low eigen values.

Chapter 4

Influence of Frost Action

The purpose of this review of frost action processes and the factors which affect frost action is to aid in the description and interpretation of lithofacies and secondary structures in Chapter 5.

Frost action is defined as the weathering process caused by repeated cycles of freezing and thawing (Washburn, 1973). A broader usage of the term would include the action of frost during freezing, thawing and subfreezing temperature fluctuations. The processes of frost action are observed where stones are ejected from fines by upfreezing. The processes are not limited to areas underlain by permafrost as is documented by the appearance of stones on previously cleared farmers' fields, the heaving of fence posts and other artificial structures.

4.1. Frost Push

Most rock types respond thermally at a different rate than soils when subjected to a temperature change. This contrast in behavior provides the basis for the frost push hypothesis. Hogbom (1910,1914), Nansen (1922), and Grawe (1936) all asserted that because the thermal conductivity of rock is greater than that of soil, the freezing isotherm should pass more quickly through stones than through surrounding soil. Hence, during downward freezing ice is thought to form at the base of a stone before freezing begins in the surrounding soil. The ice exerts a pressure, pushing the stone upwards relative to the surrounding soil. Upon thaw, the ice pedestal holds the stone up long enough for thawed material to slump around its sides, preventing the stone from returning to its original position. In a more recent statement of hypothesis, Bowley and Burghart

(1971) argued that thermal diffusivity rather than thermal conductivity was the relevant parameter. Taber (1943) rejected the Hogbom hypothesis for the upfreezing of buried stones because of the difficulty of pushing a stone into overlying already frozen material. He concludes that stones at or near the surface may undergo a frost-push-like process, but argues that there will be no net motion at the completion of the thaw. Despite this serious objection and a lack of experimental proof, the frost-push hypothesis continues to be cited in the literature.

4.2. Frost Pull

Hamberg (1915) first proposed the frost-pull hypothesis. As freezing progresses, the upper part of the stone freezes firmly into the soil, but its lower part remains relatively unconfined. When coupling with the frozen soil is strong enough, the stone moves upwards as the soil heaves. During thaw, the lower portion of the stone is anchored in the frozen soil, while the thawed soil above it settles and slumps around it. This hypothesis was perhaps best outlined in Beskow's (1930) cartoon which depicts ice lenses in the heaving soil, and a cavity opening below the stone during the freeze. He suggested that the cavity may narrow by lateral frost thrusting during freezing as well as slumping during thaw. The most convincing support of the frost-pull mechanism so far is a time-lapse movie of an upfreezing experiment made by Kaplar (1965). This side view of a spherical clast in a frost susceptible soil recalls Beskow's cartoon, but differs from it in a few details. A gap formed above the clast before it began to heave. The cavity beneath the clast filled with water, which subsequently froze. This ice filled cavity below the clast is reminiscent of the ice pedestal of the frost-push hypothesis, but it appears to form as a result of rather than a cause of the clast rotation. Several experiments

implicitly support the frost-pull hypothesis. Burrous (1978), experimenting with glass objects buried in silt, observed that the objects appeared to move upward only during the thaw phase. During the freeze cycles, they did not break through the surface of the silt or increase their protrusion above the surface. Johnson and Hansen (1974) also observed that dowels moved relative to the soil surface only during thaw. These observations are consistent with the mechanics of frost-pull: motion relative to the surface occurs only during the thaw phase. This contrasts with the frost-push, where the object is inferred to more upward relative to the surface during both freezing and thawing. Both Burrous (1978) and Johnson and Hansen (1974) found that the projection of the height of an object into a vertical axis, or its "effective height", is more important than the actual length of an object in determining its heave rate. Burrows in particular found that the effective height was well correlated with the net heave rate. Factors that maximize ice lens growth also favor high rates of upfreezing. For instance, Johnson and Hansen (1974) showed that high moisture contents and slow frost penetration increased both frost heaving in the soil and upward movement of dowels.

4.3. Inclined Freezing Front

In regions underlain by permafrost, freezing can proceed upward from the base of the active layer as well as downward from the surface (Mackay, 1983). Moreover, lateral variations in vegetation topography, snow cover, moisture and grain size can alter the thermal properties of the ground, resulting in a non-horizontal (inclined) freezing front. In these situations, one cannot predict the motion of a clast by simply reversing the direction of motion in the upfreezing model for downward freezing, as suggested by Inglis (1965). Although movement during freezing usually

parallels the heat flow direction, gravitational settling during thaw will always be biased downward, regardless of freezing direction.

Experimental work on the effect of inclined freezing fronts on upfreezing of stones is sparse. Muir (1983) found considerable migration of marked stones buried in silt after a series of freeze/thaw cycles with inclined freezing planes. The ratio of the vertical and horizontal components of the clast motions were smaller than that of the heat flow. Thus the clasts followed paths much closer to horizontal than the heat flow orientations during freezing, but in the same general direction. Vilborg's (1955) field observations also suggest a horizontal component to clast movement when the freezing front is inclined. In both of these studies, the inclination of the freezing plane was a result of surface topography, so gravity-induced displacement may have contributed to clast movement.

After conducting numerous experiments with both vertically and horizontally oriented freezing fronts in soils, Corte (1962a,1962b,1963) concluded that the fine particles(<0.074mm) move away from the cooling front, opposite the direction of heat flow. Downward freezing and thawing resulted in a net downward migration of silt particles in a cohesionless, well graded sandy gravel, an effect he termed "vertical sorting". When he reversed the freezing direction, he found a downward migration of silt in the unfrozen soil above the advancing front. He called this process "mechanical sorting", and attributed it to a "shaking" of the soil above the freeze/thaw plane. A vertical freeze/thaw plane produced a net diagonally-downward migration of the fine material in front of the freezing interface and downward. Corte (1962a) cited several field examples of sorting in gravel rich materials, and

implied that the sorting resulted from the same process he observed in his laboratory experiments.

Prestrud (1987) concluded that the mechanism of upfreezing clasts during the downward freezing experiments was clearly frost-pull. A nearly horizontal freezing front propagated downward from the soil surface heaving the soil surface through the growth of ice lenses fed by water withdrawn from lower in the soil profile. The clast initially moved downward, presumably reflecting a slight compaction of the soil beneath it due to removal of this water. Upward motion of the clast began, at the same rate as the surface, when a sufficiently strong adfreeze bond formed between it and the surrounding, heaving soil. It is apparent that cavities must have opened both above and below the clast. Subsequent particle refilling of the lower cavity with soil is necessary for displacement of the clast to persist.

4.4. Reorientation Process

Based on the relative displacement during freezing, the maximum distance an object could heave in one cycle would be approximately (Kaplar, 1969, p.36):

$$Dv = \frac{Hr}{Rf} Le \qquad \text{(Equation 1)}$$

where: Dv = vertical heave of the object in one cycle (cm)
Hr = heave rate of soil (cm/day)
Rf = rate of frost penetration (cm/day)
Le = vertical dimension of a stone below its greatest horizontal diameter (cm).

This equation assumes that the object will not sink back into a cavity and thus has the maximum possible heave in one cycle. The dimensionless ratio Hr/Rf has been called the heaving ratio (Higashi, 1958). The heaving ratio is the amount of soil heave per unit depth of frost penetration, and is a measure of soil behavior during freezing. In most cases the heaving ratio increases in value with increasing depth of frost penetration. It is not uncommon in laboratory tests to find that the heave rate of the soil is relatively constant, but that the frost penetration rate shows a pronounced decrease with increasing depth of frost penetration, in rough agreement with Stephan's solution (Penner and Ueda, 1977). Stephan's solution predicts that the frost penetration rate is inversely proportional to the depth of frost penetration. For this rather common condition, the heaving ratio of the soil (Hr/Rf) will increase with increasing depth of frost penetration. Applying this result to Equation 1, this predicts that the heave of objects should increase with increasing depth of burial below the surface. For this to occur, the frost table must be below the bottom of the objects.

The parameter Le in Equation 1 is the effective height, the height of the object's orthogonal projection onto a vertical plane. Kaplar's (1969) observations on the upheaving of stones imply that the effective height of an object can be approximated by that portion of the vertical dimension below the greatest horizontal diameter.

Johnson, Muhs, and Barnhardt (1977) studied the upheave of wooden dowels that were buried with their long axes at various angles to the horizontal and their tops just touching the soil surface. Their results show that the upheave rate increased with increasing projected height (vertical dimension) of the dowels regardless of their angle. They suggest that the use of the projected height (vertical dimension) as a parameter compensates for the object's orientation. The projected height of an inclined tabular object or rod is calculated by:

$$Lv = L \sin f + t \cos f \qquad \text{(Equation 2)}$$

where Lv is the vertical dimension or projected height (cm), L is the actual length of the tabular object or rod (cm), t is its thickness (cm), and f is the angle of inclination as measured from the horizontal plane (degrees). If the object is thin (t is small) then the projected height is nearly proportional to the sine of the angle of inclination. Johnson, Muhs and Barnhardt (1977) also found that the inclined wooden dowels tend to rotate to the vertical, with an average value of 10.8° in 3 freeze-thaw cycles. This behavior is in agreement with field observations that stones in cold regions tend to be oriented with their long axes perpendicular to the surface (Washburn, 1973). One theory to explain this rotation towards the vertical makes use of the frost-pull concept. If the upper part of an inclined object is being pulled upward by the frozen soil, the resisting force of the uniform soil will be downward against the inclined surface of the object and not colinear with the heaving force (Washburn, 1980). The

result may be a force couple that rotates the object towards the vertical.

Pissart (1973) attempted to determine this experimentally. He placed thin metal rods, inclined at 45°, partially submerged in mud. The rods were subjected to slow freezing which caused migration towards the surface. This uplift was accompanied by rotational movement. Pissart (1973) concluded that the rotational movement was due to different forces acting upon the rods. He found that permafrost was not essential to rotate the objects provided that sufficient moisture was present and the material was suitable for segregated ice lenses to form.

Burrous (1978) conducted freeze-thaw experiments on objects of different shapes. He found that all originally inclined objects emerged from the soil at the same angle as they were emplaced, and that most rotation occurred after emergence. He also found that many of the originally inclined objects were oriented horizontally after emergence. He believed that this rotation occurred during thaw. During thaw, previously expanded soil must contract vertically as the segregated ice melts and the excess water is expelled. Rigid objects cannot contract and thus the contracting soil must either flow around them or rotate the objects towards the horizontal. In any case, there must be a higher pressure on the upper side of an inclined object than on the lower side. When the thaw front first encounters an object, the object is still held as a rigid cantilever by the soil still frozen around its lower end, and so the thawed, consolidating soil must flow around the object. The zero isotherm can penetrate more quickly through the stone than through the soil. This is partially because the thermal diffusivity of clasts is larger than that of the thawed or frozen soil and partially because the effective diffusivity of the soil becomes very low at the thaw or freezing front due to water's high latent heat of fusion. A thin layer of soil will thaw around the lower part of the object before

the general thaw front advances past the bottom of the object. The inclined object will then rotate to relieve the higher force on the upper side of its slanted surface.

4.5. Factors affecting the growth of ice lenses

The freezing of frost-susceptible soil with water available normally involves opposing actions: the downward advance of the freezing front and upward frost heave. Heaving is the result of ice segregation during the freezing process. The advance of the freezing front causes alternating bands of soil and ice to form. The external manifestation is frost heave (Prestrud, 1989). As well as frost heave, the vertical migration of objects is generally attributed to the formation of ice lenses during freezing. For this to happen, it is generally agreed that: 1) subfreezing temperatures; 2) water; and 3) a frost-susceptible soil must be present. With all of these factors present the degree of frost-susceptibility may vary with the rate of heat removal, the temperature gradient, the mobility of the water, the depth to the water table, the overburden stress, the soil density and the soil texture.

The most important factors affecting frost susceptibility are: 1) soil texture; 2)pore size; 3) mineralogy; 4) rate of heat removal; 5) temperature gradient; 6) moisture conditions; 7) overburden stress or surcharge; and 8) freeze thaw cycling.

4.5.1. Soil Texture

Grain size is used as the basis for most frost-susceptibility criteria because it is the most easily measured soil property. Taber (1929) was the first to demonstrate that surface uplift following freezing from the surface downward of certain materials containing water was not the result of expansion of water cooling to form ice. Taber's (1929, 1930) experiments showed that simple freezing of interstitial water causes practically no uplift. His

experiments also showed that frost heaving occurred readily on silts, less readily on clays, but never on sands. Examination of the frost heaved materials revealed that uplift was caused by the formation of segregated ice within the materials. Ice segregation only occurs in materials (termed frost susceptible) that have a suitable grain size composition.

Casagrande (1931) found that under natural freezing conditions and with sufficient water supply, considerable ice segregation develops in poorly sorted soils containing more than 3% grains smaller than 0.02 mm, and in very well sorted soils containing more than 10% grains smaller than 0.02 mm.

Leary et al (1968) concluded that the grain size effect is very complex. Only a certain fraction of particle sizes in a soil influences frost heave behavior, and the amount and activity of the clay-sized particles and the uniformity of the gradation of soil particles less than 74μ in diameter are controlling factors.

Many authors have classified gravels as having slight to no frost susceptibility (Ruckli, 1950; Bonnard and Recordon, 1958; Jessberger, 1976). Ruckli (1950) based his classification on grain size. Bonnard and Recordon (1958) considered soils to be frost susceptible if more than 3% of the soil particles were less than 0.02 mm in diameter generally agreeing with Casagrande's (1931) criteria.

Pewe et al (1969) were among the first to consider gravels as frost-susceptible. They examined glacial outwash deposits, classifying them as unmodified or those that had been modified (and thus frost susceptible). They found that the ranges of occurrence of the cumulative frequency curves for both modified and unmodified outwash have the same general form but much greater variability in the size of the modified outwash is indicated. The modified outwash is generally less well sorted and more positively

skewed, indicating a greater amount of fine material in an essentially coarse grained distribution.

Brandl (1970) conducted freezing tests on gravels from which he concluded that loamy gravel material with a percentage of fines of about 8% must be regarded as frost susceptible. He found that the amount of clasts was not as critical as the amount and size distribution of the matrix providing that the mixture is matrix-supported.

4.5.2 Pore size

Pore size and therefore grain size strongly affect the growth and form of ice in the soil by influencing: 1) the freezing temperature of the phase-boundary water and the water in fine capillaries; and 2) the movement of the water to the freezing front. It is well known that the amount of unfrozen water varies inversely with the grain size, other conditions remaining equal. This accounts for the fact that silt and particularly clay freeze at a lower temperature than coarser material. The basic factor accounting for the different freezing temperatures is the greatly increased contact area between solids and water as grain size is reduced, and the resulting increased tendency for surface effects to keep capillary and phase-boundary water from crystallizing. Thus, for a closed system the finer the grain size and the higher the freezing temperature, the less the amount of ice. In an open system where water is continuously available for movement to the freezing plane, the water in material of fine grain size may remain mobile longer and hence larger ice lenses may build up. Grain size influences movement of water to the freezing front because the potential for drawing water to the freezing front increases with decreasing grain size. This suction potential is absent in coarse sand and gravels lacking fines, so that the segregation of ice lenses opposed to the development of interstitial ice is inhibited. Soils may be so fine grained as to become essentially impermeable.

Consequently, the movement of water to the freezing front to form ice lenses and cause heaving is at a maximum where the suction potential and permeability combine most effectively. Silt is particularly prone to heaving, as it permits relatively easy migration of moisture during freezing, and in an open system this factor greatly favors the development of greater ice content in silt than in clay. A soil is considered frost susceptible if it has 1) several percent of particles less than 0.07 mm in diameter; or 2) contains more than 3% to 10% particles less than 0.02 mm in diameter.

4.5.3 Mineralogy

The mineralogy of fines can influence the freezing process with respect to the migration of water and heaving. Clays with expandable structures are able to hold more water but the water is relatively immobile compared to that in non-expandable clays. Strong frost-heaving is more likely to be associated with kaolinite than with bentonite or montmorillonite (Grim, 1952; Linell and Kaplar, 1959). Salts such as NaCl and CaCl2 influence freezing by depressing the freezing point significantly (Yoder, 1955) and have also been found to be high heave producers (Lambe et al, 1969).

For silty-sand deposits which do not contain carbonates and have varying amounts and combinations of illite, vermiculite, and chlorite, the relative frost susceptibility depends approximately upon the amount of fines (less than 0.02 mm) present (Lambe et al, 1969). Minerals having the highest frost heave producing ability are, in increasing order: muscovite, calcite, magnesite, iron montmorillonite, illite, kaolinite, nontronite, and attapulgite (Lambe et al, 1969).

4.5.4. Overburden pressure

In a study of material from Beaver Creek and White River in the Yukon, Hardy Associates (1979) found that the frost

susceptibility of gravels was greatly reduced at pressures of 100kPa, which corresponds to depth of overburden of 4 to 7 m depending on the position of the water table. The Beaver Creek materials had a larger percentage of fines than the White River soil, but a lower percentage of clay-sized particles. Overburden pressure was less effective in suppressing heave in very fine grained soils than in coarser grained soils.

4.5.5 Clast Shape

Three large scale tests of frost action within the active layer were conducted by Burrous (1978). Three different shapes, laths, discs, and spheres, with two different sizes of each were placed at various depths within a sample of the Fairbanks silt. The laths and discs were placed at different angles of inclination. Burrous (1978) found that the laths were heaved faster than the discs which in turn were heaved faster than the spheres. This is expected given the projected heights of each object. Despite the differing heave rates, only the ratio for laths to spheres was statistically significant. Burrous (1978) concluded that the shapes chosen did not have much influence on heave rate.

Chapter 5

Descriptions and Interpretations

5.1. Lithofacies Descriptions and Interpretations

The term "facies" will be used in the modern sense given by DeRaaf et al. (1965) who distinguished sedimentary facies on the basis of lithological, structural, and organic aspects visible in the field. Because of the absence of biogenic structures in the sections studied, throughout the rest of the text the term "facies" will be used synonymously with "lithofacies".

Facies were distinguished on the basis of 1) primary sedimentary structures, including trough cross-bedding, and horizontal bedding, and 2) textures, including clast-supported and open-worked, clast-supported and matrix-filled or matrix-supported gravels. Secondary sedimentary structures were not used to distinguish facies and will be described separately. This was done in an attempt to determine if primary characteristics influence the formation of secondary structures. Although description and interpretation of primary lithofacies is not an objective of the study, it is inherently necessary to determine the result of primary deposition before we can elucidate on secondary processes.

Eleven facies were recognized in the study area. Of these eleven facies, five were common and found in numerous sections in a repetitive sequence, whereas others were restricted to one or two sections. The most common facies, lithofacies 1 through 5, will be described and interpreted before the less common facies are discussed. To enable the reader to form a clearer image of the associations between the facies and their areal distribution, Figure 5.1 shows the location of the sections in the study area. A detailed discussion of the facies associations, the common sequences

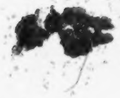


FIGURE 5.1 SECTION LOCATIONS

0 10 25 50
km

Cypress Hills and Wood Mountain Formations

Alberta

Saskatchewan

Maple Creek

Kealey Springs

Police Point

Center Block

Ravenscrag

Horseshoe Canyon

Cypress Hills

Frenchman River

Assiniboia

Wood Mountain

Cedar Creek

Wideview

Resch

Rockglen

St. John's

United States

of facies and their distribution through the study area will be undertaken in Chapter 6.

Lithofacies 1: Horizontally stratified, pebble-matrix supported gravel

Description:

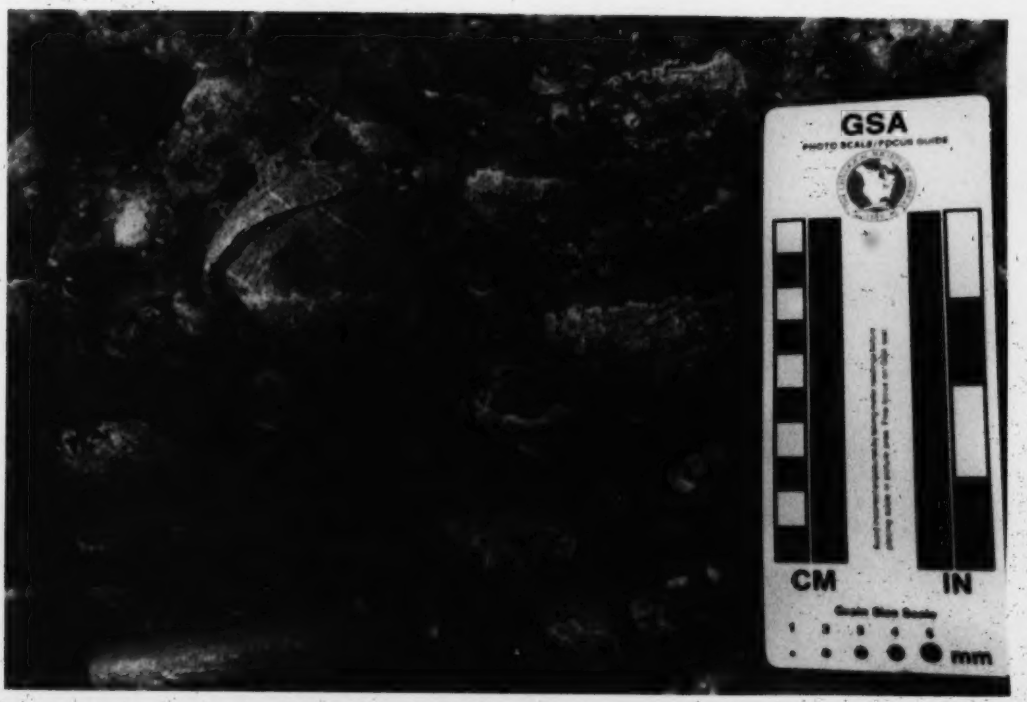
This facies is composed of approximately 20% well sorted large pebbles and cobbles floating in a "matrix" of fine pebbles (Plate 5.1). The clasts are elongated to discose, well-rounded, and either display a weak imbrication or are oriented parallel to the bedding planes with the intermediate (b) axes dipping as steeply as 20°. The fine pebbles in the matrix, comprising up to 60% of the facies, are equant, moderately rounded and well-sorted. A secondary infilling of silty sand is present in quantities up to 20%. The facies displays an indistinct planar stratification and crudely developed horizontal bedding with individual beds 2 to 15 cm thick. Basal contacts are planar, distinct and on a small scale undulatory over 10 cm. Lateral extent of individual units is up to 140 m. Commonly this facies contains a secondary carbonate deposit along the underside of the clasts.

Interpretation:

The coarseness, sorting, and structure of Lithofacies 1 suggest an active bedload system and relatively high sediment concentration in flows. The pebble matrix-supported framework indicates that currents were capable of winnowing finer grained sediments. A fluvial origin for this facies is supported by the presence of imbricated clasts with intermediate axis (b) imbricated and long axis (a) transverse to flow. Grain size differentiation on a bar surface followed by progradation of lateral movement may be the cause where well-sorted, coarse pebble beds with a fine-

Plate 5.1

Typical exposure of lithofacies 1



pebble matrix do not show any clear grading. In Hein and Walker's (1977) model, high sediment and water discharges promote downstream growth of horizontally stratified gravel sheets. With continued vertical and lateral growth the sheets develop into longitudinal and/or diagonal bars. A similar result may be achieved by a falling water state (Steel and Thompson, 1983). Similar interpretations for this facies are reported from proximal reaches of modern braided outwash streams (Boothroyd and Ashley, 1975; Miall, 1977) and from ancient braidplain deposits (Eriksson and Vos, 1979; Middleton and Trujillo,1984).

Lithofacies 2: Structureless, matrix supported gravel

Description:

This facies consists primarily of poorly sorted,cobble sized clasts although fine pebbles and small boulders are also present. Clasts are angular to subrounded, with angularity increasing with decreasing clast size. The matrix is dominantly sand with some silt and clay. Clast fabric is commonly randomly oriented. Although clast imbrication is rare, some elongate clasts are oriented with their long axes horizontal to subhorizontal and intermediate axes inclined. Sedimentary structures are absent and distinct bedding planes are difficult to recognize. Grading is rare although both normal and inverse coarse-tail graded sequences do occur. Basal contacts are sharp, and flat to irregularly scoured. Units within this facies tend to be tabular and can be traced 50 to 100 m. Post-depositional modifications to this lithofacies include a well developed re-orientation of clasts into a vertical/subvertical orientation.

Interpretation:

Important characteristics of this lithofacies with respect to depositional processes include the coarse-grained texture and angularity of many clasts, poor sorting, sandy

and/or muddy matrix support, randomly oriented fabric, and absence of sedimentary structures. Although the coarseness of the clasts suggests high energy flow conditions, the textural immaturity of the unit (poor sorting, high matrix content) indicates that few grain to grain bedload collisions occurred and the transporting medium was incapable of winnowing fines and sorting the clasts. Therefore, both the matrix and the clasts were transported and deposited together.

High sediment concentration flows such as debris flows on alluvial fans are characterized by many of the features described above (Bull, 1964, 1972). The ability of debris flows to move on relatively gentle slopes and with a minimum of very fine matrix (clay and silt) has been documented experimentally (Rodine and Johnson, 1976). High viscous debris flows are characterized by a muddy matrix and commonly occur in the proximal reaches of alluvial fans whereas sandy matrices are typical of less viscous, distal fan debris flows.

The large basal clast dispersions are maintained by high dispersive pressures, as indicated by beds possessing a strong a-axis parallel and imbricated upstream. Some beds show a more randomly oriented fabric however. Variation in clast fabrics is probably related to the degree of grain mobility associated with the distance of flow. Well developed fabrics originate from flows which have had sufficient time for dispersive pressure effects to orient the clasts. Poorly-developed fabric is a consequence of insufficient time to create an organized fabric because of a short travel distance. Poor sorting and absence of grading within the beds implies limited grain mobility during flow, a result of suppression of fluid turbulence. This feature may also be explained by rapid deposition from poorly-stratified turbulent flows however.

Dispersed pebbly sands are deposited from "pure" grain flows. Parallel orientation and imbrication of coarse

clasts indicate transport and deposition from flows which were dominated by dispersive pressure. With the availability of high slopes, these beds are set into motion by the development of dispersive pressure. En masse deposition occurs because of the decreasing flow competence related to decreasing shear stress. This condition is probably the result of decreased regional slope and insufficient development of a secondary supporting mechanism. Irregular basal scour illustrates the fluidity and high energy of these flows.

Lithofacies 3: Horizontally stratified, matrix supported gravel

Description:

This facies is composed of single beds 15 to 40 cm thick. The clasts are well-sorted, coarse pebbles to fine cobbles. Clasts tend to be elongated, discose and well rounded. The matrix is composed of well sorted coarse-grained sand with trace amounts of silt and clay. Clast fabric is organized with long axes parallel to bedding and rare imbrication. Basal contacts are distinct and broadly planar. Units extend laterally for up to 140 m.

Interpretation:

The sand matrix cannot be considered to result from late-stage infiltration because it supports the clasts. If the coarse pebbles/fine cobbles had been transported and deposited by rolling on the bed, then currents should easily have been capable of causing the sand to bypass. Steel and Thompson (1983) interpret these cases as where the sand and clasts were deposited more or less simultaneously, and from a flow in which there was high sediment concentration and rapid deposition. The distinctive bimodality implies that previously sorted gravel was entrained and redeposited during flood.

Lithofacies 4: Structureless, clast supported gravel

Description:

Clasts within this facies are moderately sorted, coarse pebbles. Clasts are generally elongated to discose and well rounded. The matrix is largely poorly sorted silty-sand, with no apparent stratification although normal grading is rarely present. The clasts display a random orientation to a weak preferred fabric. Imbrication was not observed. The beds are laterally continuous for up to 100 m and the basal contacts are distinct, flat to scoured.

Interpretation:

Facies characteristics such as very poor sorting, lack of stratification, crude normal grading, and non-erosive bases suggest sedimentation through mass emplacement (Harms et al, 1982; Nemec and Steel, 1984; Krapez, 1985). Normal grading is developed as denser particles settle directly from suspension, whereas modal gravel clasts are emplaced at higher water levels which are achieved during the flow event (Nemec and Steel, 1984). This forms a basal lag, and results in an overall normal grading of gravel clasts in the sediment.

Lithofacies 5: Crudely stratified, clast supported gravel

Description:

This facies is dominated by pebbles that display well developed imbrication with A-axes transverse to the dipping A/B plane. The matrix is a poorly sorted silty sand in quantities up to 25%. Stratification is planar, 0.3 to 1.5 m thick, and is discernable through changes in gravel texture. The beds are internally ungraded. Typical stratification includes thinner small pebble gravelly beds, thicker pebble to cobble gravel beds, and discontinuous

sandy lenticular beds. Beds are laterally continuous for up to 100 m and the basal contacts are flat to slightly scoured and distinct.

Interpretation:

Facies characteristics such as poor sorting, coarse texture, dominant a-axis flow parallel and a-axis imbrication, and scattered fabric patterns suggest deposition as gravelly sediment flows. However, several workers have described fluvial deposits such as braid bars to be internally massive to crudely planar stratified (Enyon and Walker, 1974; Miall, 1977,1978; Hein, 1984; Smith, 1985). Also, low relief channel features such as diffuse gravel sheets may develop crude horizontal stratification (Hein and Walker, 1977; Hein, 1984). Low relief channel features and braid bar deposits should have distinct a-axis flow transverse, b-axis imbrication upstream if the gravelly sediment is coarse and transported as bedload (Hein, 1984). Only saltation and pivoting of smaller clasts in channels will produce a-axis flow parallel, a-axis imbricate patterns (Hein, 1984). Thus, development of a-axis imbrication in stream deposited crudely stratified gravel is possible only if current velocities have the ability to rework bedload sediment. Interpretation of this facies based only on sedimentary structures is difficult. A proper interpretation should be based upon numerous characteristics, most of which (such as associations) have not been described in this study.

Lithofacies 6: Trough cross-stratified, clast supported gravel

Description:

Cross-bedded sets of this facies range from 15 to 40 cm thick with most less than 25 cm, and consist of clast-supported pebble gravels. Clasts are moderately to poorly

sorted, elongate, and subrounded to rounded. The matrix is silty sand. These units occur either as single sets or in cosets of up to six sets. The beds are laterally discontinuous and display lenticular geometries of several meters long and only a few meters wide, aligned parallel to flow. The basal contacts are sharp and scoured.

Interpretation:

This facies represents channel deposits which formed as a result of migrating large-scale three dimensional ripples or dunes. This type of tractional bedload transport requires competent flow and high flow velocities (Harms et al, 1975,1982). Cross-stratified gravels were interpreted by Leckie and Cheel (1989) to represent local cross-channel bars. The limited lateral extent of the cross-stratal sets and their association with imbricate gravels suggests that the cross-bedded gravel formed on the downstream edges of longitudinal bars (Rust and Koster, 1984).

Lithofacies 7: Openwork gravel

Description:

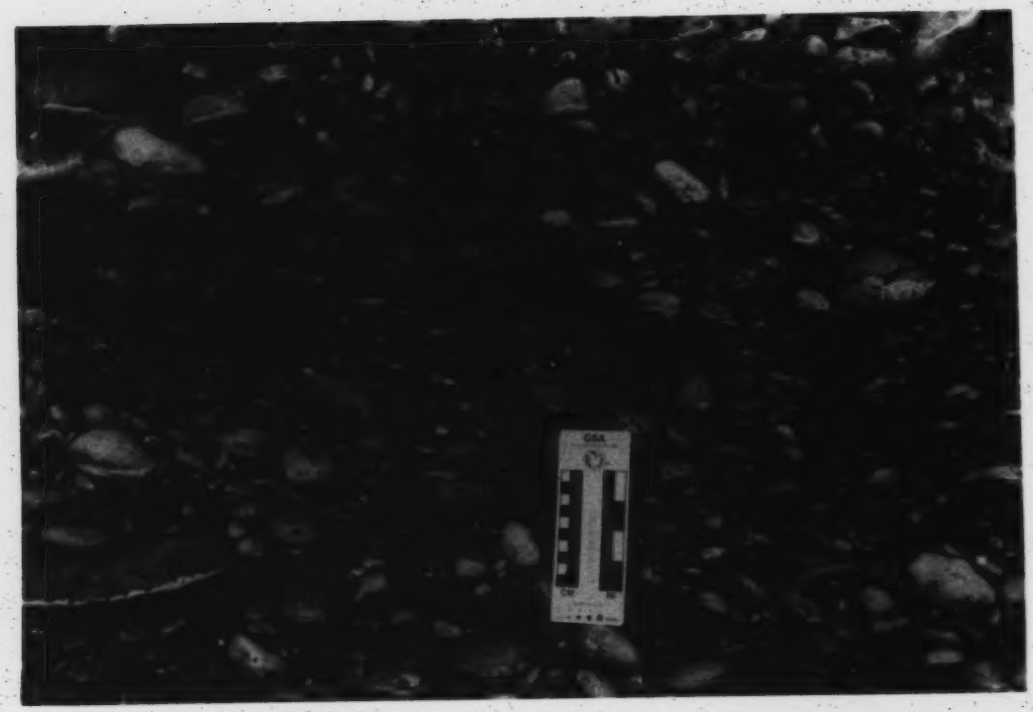
This facies is comprised of moderately to well sorted, well rounded, equant to elongate pebbles. The facies is structureless (Plate 5.2). The clasts display a random orientation. Some matrix, never more than 10%, is present as a secondary infilling. Beds are laterally discontinuous, biconvex to convexo-planar lenses up to 3.2 m long and 0.1 to 0.8 m thick.

Interpretation:

Lithofacies 7 is the result of gravelly bedform sedimentation under high discharge conditions. Openwork gravel represents fluctuating discharge levels with current velocities high enough to maintain finer sediment in

Plate 5.2

Typical exposure of lithofacies 7 in lens outlined



suspension, assuming that finer sediment is available (Smith, 1985). Small areas of matrix-filled gravel within the openwork gravel indicate either sporadic discharge conditions or clogging of gravel pockets during aggradation under decreased flow conditions (Frostick et al, 1984).

Lithofacies 8: Massive to horizontally laminated sands

Description:

This facies is composed of well-sorted coarse sand to moderately-sorted clayey-coarse sand. Laminations are 0.3 to 2 mm thick and laterally discontinuous over 10 cm. Some normal grading is present. At some locations, the laminations are convoluted. The basal contacts are sharp, trough to planar. This facies occurs 5 to 30 cm thick and is laterally continuous for up to 24 m. At some locations where the laminations are contorted, the facies appear to pinch and swell.

Interpretation:

The unstratified beds of well sorted coarse sand interpreted to reflect rapid deposition from suspension. In the case of laminated beds, deposition resulted from lower velocity waning discharge within channels or on bartops. Experimental flume work reported by Middleton and Southard (1984) suggests that at most naturally-occurring mean flow velocities and depths, deposition of this size sediment results in laterally extensive massive sheets or beds.

Lithofacies 9: Type B Climbing Rippled Sands

Description:

This facies is composed of well-sorted medium to coarse grained sands. The ripples present are sigmoidal shaped. Laminations are 0.2 to 0.3 cm thick. The elevation differences between the crest and the trough of individual

ripples appear subdued. The ripples climb at an angle of approximately 40° and grade laterally into massive sands.

Interpretation:

Climbing ripples are formed when the rate of addition of new sediment to the bed is greater than the rate at which sediment is being transported from the stoss to the lee side of the ripple (Harms et al, 1982). Type B, or supercritical climbing ripples are those with angles of climb sufficient for preservation of the full ripple profile. Type B climbing ripples with an angle of climb greater than the slope angles on the stoss sides of the ripples are formed by abundant fallout and very slow ripple migration (Harms et al, 1982).

Lithofacies 10: Trough Cross-bedded Sands

Description:

Moderately sorted, coarse to fine grained sands comprise the bulk of this facies with rare pebbles occurring along the base of some troughs. The troughs are 20 cm deep and up to 3 m in length. Individual sets exhibit normal grading. Basal contacts are scoured.

Interpretation:

Well-sorted, trough cross-bedded sands are considered to reflect deposition from large-scale three-dimensional ripples and dunes (Middleton and Southard, 1984). Based on data presented in bedform stability plots, Blatt et al, (1980) suggested ripples of this size tend to be flattened at higher velocities. Those bedforms examined in the field had normal geometries, and based on flow depths and velocities from reported flume studies (Middleton and Southard, 1984), it is suggested that most of these bedforms developed in shallow channels with moderate flow velocities.

Lithofacies 11: Massive silts with rare pebbles

Description:

This facies is composed of well-sorted silt with minor amounts of clay to fine-grained sand present. Clasts occur in quantities of up to 5% and have a dominantly random orientation. The silts do not display any sedimentary structures. The facies ranges in thickness from 5 to 35 cm. The basal contacts are distinct, undulatory and conformable. This facies occurs as the uppermost exposed facies, when present.

Interpretation:

Massive well-sorted silt represents rapid deposition of suspended sediment by aeolian processes. The rare clasts are most likely the result of freeze-thaw processes, lifting the clasts from underlying gravel facies.

Chapter 6

Descriptions and Interpretations of secondary structures

Three types of secondary structures, a)contortions, b) pillars, and c) reoriented units, have been observed in the study area. The contortions are believed to be secondary structures as they typically involve the contacts between units. The pillars are large vertical structures which cross-cut bedding, and the reoriented units display a strong vertical clast fabric.

6.1. Contortions

Contortions are defined in this study as those units which display highly irregular bedding surfaces (Plate 5.3).

6.1.1.Descriptions

Resch Site

The Resch site is located on the uppermost bench of the Wood Mountain uplands near Macworth, Saskatchewan, at an elevation of approximately 1100 masl (Figure 5.1). The lowermost exposed facies observed is lithofacies 5, crudely stratified clast-supported gravel (Plate 5.4). The facies is up to 65 cm thick and contains a convexo-planar lens of lithofacies 8. This lithofacies does not display any evidence of disturbance. This lithofacies is overlain by a poorly sorted matrix supported gravel (lithofacies 2) containing pockets and injections of the overlying massive clast supported gravel (lithofacies 4). The contact between lithofacies 2 and 4 is sharp,highly irregular and contorted (Plate 5.5). The pocket is 40 cm wide and 30 cm high. Maximum clast intermediate (b) axis length is 12 cm. The

Plate 5.3

Typical exposure of contortions (clast supported gravel facies overlying a matrix supported facies)

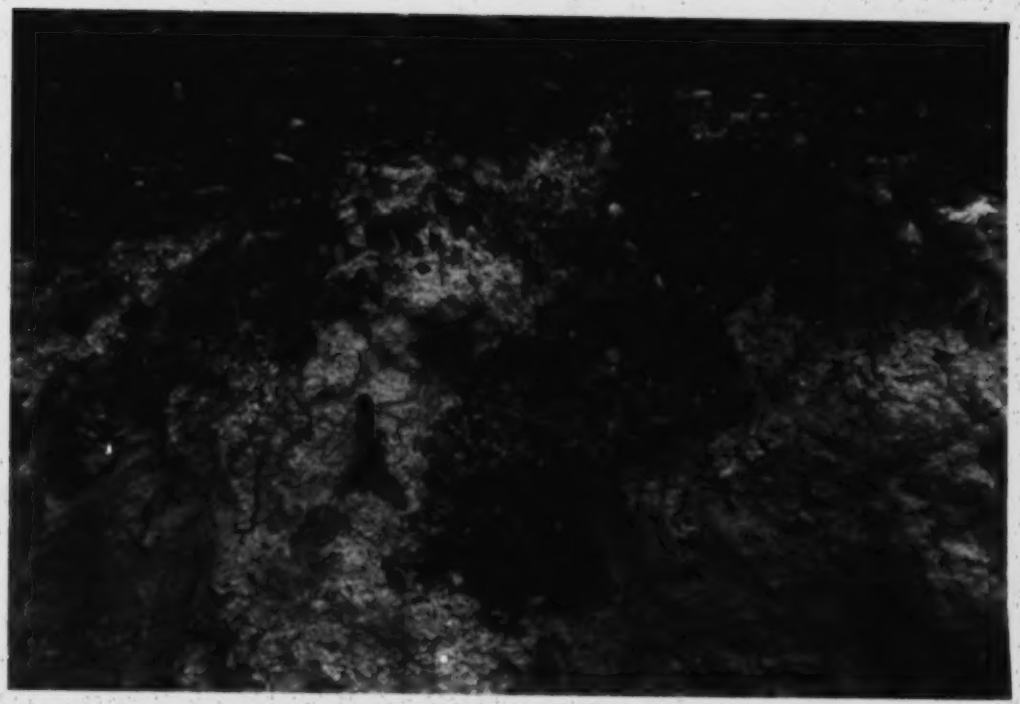


Plate 5.4

Panoramic view of Resch Site showing contortions

Plate 5.5

Close up of the contortions at the Resch Site

pebbly-cobble pocket contains up to 25% silty sand as a matrix. The clasts in the pocket display a weak to moderate preferred subvertical orientation. Surrounding the pocket, lithofacies 2 displays a wide variety of clast percentages, varying from 5% to 25%. The matrix is poorly sorted and contains a small (<3%) amount of carbonates. This facies varies in thickness from 20 to 75 cm. The top 25 cm of the exposed section is composed of lithofacies 11, massive silt with rare pebbles.

Center Block Site

The Center Block section is a gravel pit within Saskatchewan Cypress Hills Provincial Park at an elevation of 1375 masl (Figure 5.1). The lowermost exposed lithofacies is horizontally stratified, matrix-supported gravel (lithofacies 3) (Plate 5.6). The basal contact of this unit is covered and the facies is up to 70 cm thick. This lithofacies grades vertically into structureless, matrix-supported gravel. The observed change is due to a loss of stratification which may be the result of a change in clast orientation. In the massive portion of the matrix-supported gravel, the average clast dip is 82.6° whereas in the stratified portion, the observed dip is 21.2°. This lithofacies is irregularly overlain by lithofacies 4, a massive clast supported gravel. Also, within this lithofacies, the apparent massive nature is enhanced by the near-vertical dip of the clasts. These convoluted units are unconformably overlain by lithofacies 6 (trough cross-stratified gravel). The basal contact is sharp, trough shaped, and truncates the contortions of the underlying lithofacies. This unconformity is evidenced by the presence of shield derived clasts within lithofacies 6 and the lack of such clasts below the contact.

Plate 5.6

Contortions at the Center Block Site

Ravenscrag Site

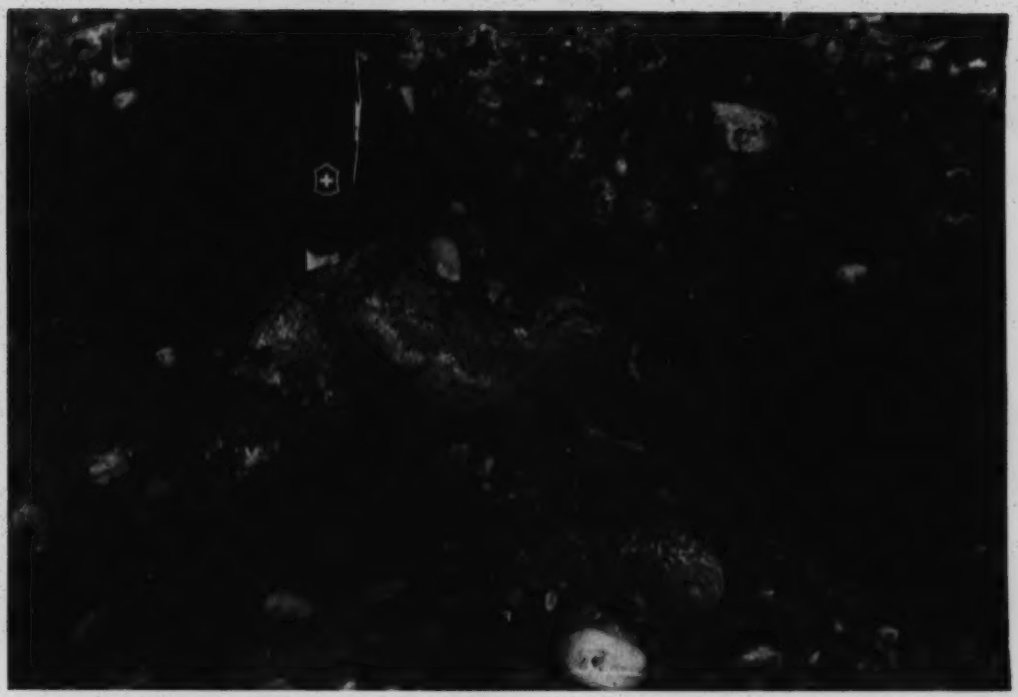
Approximately six kilometers north of the hamlet of Ravenscrag, at approximately 1300 masl, is the Ravenscrag gravel pit. The Ravenscrag section is composed of interbeds of lithofacies 3 and lithofacies 5. Lithofacies 5 is crudely stratified, clast-supported gravel, 5-12 cm thick, and it alternates with 25-30 cm thick beds of horizontally stratified, pebble-matrix supported gravels. These units are laterally continuous throughout the entire exposure, approximately 45 m. These tabular units are conformably overlain by a 3-30 cm thick exposure of lithofacies 8, laminated to massive sands. These clayey-sands are finely laminated but extremely contorted. Laminations can be traced for up to 10 cm laterally. The laminations appear to thicken where the bed thickens. Laminations were observed to deform beneath clasts (Plate 5.7). The deformed zone extends for up to 12 cm (the thickness of the bed) beneath these clasts. In general, the thickness of the bed appears to thicken to the west. These laminated sands are overlain by lithofacies 2, structureless matrix supported gravel. The contact between these units is sharp and highly irregular. The contact is also marked by a thin, discontinuous layer, up to 2 cm thick, of small calcareous nodules, 2 mm in diameter, well-rounded and moderately sorted. This well-cemented, calcareous nodule layer is interpreted as a Cca horizon of the modern soil. Lithofacies 2 is characterized by a poorly sorted matrix, and a strong sub-vertical orientation to the elongated clasts.

Police Point Site

The Police Point section is located on the Alberta side of the west block of the Cypress Hills at 1500 masl. It is a large, recent (within the last ten years) slump scarp which exposes 12m of sediments. The basal sediments exposed

Plate 5.7

Contortions in the sand facies at the Ravenscrag Site



in this section are white bentonitic beds probably of the Paleocene Ravenscrag Formation. These sediments are exposed to a maximum of 2m, are finely laminated and bedded.

The Paleocene Ravenscrag Formation is overlain by up to 8m of coarse Cypress Hills Formation. The basal contact is sharp and roughly planar. The exposure of Cypress Hills Formation consists of interbeds of lithofacies 4 (massive clast-supported gravel), lithofacies 5 (crudely stratified clast-supported gravel), lithofacies 6 (trough cross-stratified clast-supported gravel) and lithofacies 3 (horizontally stratified matrix-supported gravel). The lithofacies occur as lensoid interbeds to more planar tabular geometries, ranging in lateral exposure from 1.2 to 110 m. Beds are 0.25 to 2 m thick. Individual lenses contain iron and/or manganese staining. This lower portion of the Cypress Hills Formation is well cemented.

The uppermost 2.0 to 2.5 m of exposure is uncemented Cypress Hills Formation, lithofacies 2, and loess, lithofacies 11. This upper portion is contorted with sharp and highly irregular contacts (Plate 5.8). The contortions vary in shape from bulbous festoon to flame-like structures.

6.1.2. Interpretation of Contortions

Soft Sediment Deformations

Fluid movement through most grain supported sediments involves the slow percolation of water in response to broad pressure gradients. The resulting fluid drag on the sedimentary grains is generally negligible. Particle adjustments represent the slow transformation of an unstable grain framework into one of greater stability. Seepage cannot account for the formation of sedimentary structures unless flow rates become large, in which case sediment strength may be substantially reduced and a liquefied or

Plate 5.8

Contortions at the Police Point Site

fluidized state approached. Liquefaction occurs when a metastable or under-consolidated sediment experiences a sudden loss of shearing resistance associated with grain framework collapse and a rapid but temporary increase of pore-fluid pressure. Effectively, a grain-supported sediment is transformed into a transitory fluid-supported suspension. Fluidization occurs when a fluid flowing vertically through a granular aggregate exerts sufficient drag force on the grains to lift or suspend them momentarily against the force of gravity. When the full size range of detrital sediments is considered, it is apparent that full fluidization of natural sediments is likely only for coarse grained silt and sand-sized material (Lowe, 1975). Both coarse sand and gravel, and fine silt and clay dewater by seepage. Coarse grained sand and gravel is usually highly permeable. Water escapes readily through the wide pore-spaces, and the grains offer a relatively large mass resistance to lift and fluidization. Materials such as clay, sand-sized phyllosilicates, and organic grains decrease sediment permeability and promote inhomogeneous liquefaction even when present in small amounts (Lowe, 1975). A strongly bimodal sediment is likely to show fluidization of the smaller particle population within the pores of the undisturbed larger grains. Thus silt or clay may be fluidized and elutriated within the pores of coarser sand or gravel while the overall grain framework remains intact.

Many structures form in response to gravitational instabilities arising from the superposition of more dense over less dense sediments. These instabilities, generally of a Rayleigh-Taylor type (Chandrasckhar, 1961), originate both through the deposition of gravel, sand or silt on mud or clay, or through inhomogeneous liquifaction and fluidization.

Soft sediment deformations are non-tectonically induced deformations in sediments before they have been consolidated. Soft sediment deformation occurs in response to stresses originating in four main ways : 1) from gravitationally unstable vertical profiles of sediment bulk density; 2) non-uniform confining loads; 3) presence of the sediment on a slope; and 4) fluid flow (Allen, 1984). High pore-fluid pressures can build within confined sands as surrounding shales compact and become impermeable (Lowe, 1975). Where breaks develop in confining sediments by jointing, faulting or mass-movement, the rush of fluids toward the fracture give rise to fluid stresses greatly in excess of those required to fluidize the sand (Lowe, 1975).

A similar situation where confined water-saturated sediment is confined occurs in pingos. A pingo is an ice-cored hill which has been domed up from beneath by the intrusion of water under pressure (open system). Groundwater in permafrost regions is subject to artesian pressure. The role played by permafrost is to act as an impermeable and confining layer for sub-permafrost waters. Where the permafrost is thin or discontinuous, artesian pressures may develop such that the overlying permafrost is ruptured (French, 1976). Theoretical calculations by Holmes et al (1968) suggest that the total force required to maintain a pingo 30 m high (the average pingo height (Washburn, 1973)) and to overcome the tensile strength of the ground and force it upwards is considerably greater than the highest artesian pressure yet measured. Swinzow (1969) cites reports of a pingo that was under sufficient pressure to rupture explosively and throw blocks of frozen ground a distance of 15-20 m. Although the geological conditions necessary for pingo formation are not exhibited by the Cypress Hills and Wood Mountain sediments, a similar

process, such as those which form involutions, may have acted on the sediments.

Involutions

The term "involution" is used for structures consisting largely of 'deformed involved and haphazardly interpenetrating masses of silt and sand originally arranged in a horizontal position (Sharpe, 1942). In a broader meaning the term comprises a pattern of unoriented or vertical contortions which are sometimes termed 'cryoturbations' or 'frost deformations'. Involutions result from frost action and many authors believe that they require permafrost conditions. Although many authors consider involutions are due to pressure by the growth and melting of ice, no observations have been made. Exactly how periglacial involutions are formed is not established. A few authors have commented on similar or identical structures developed in marine, lacustrine, deltaic and fluvial deposits. Both the periglacial and non-periglacial deformations are closely linked, as each requires a boundary surface between sediments differing in mechanical properties. The structures result from the tendency to restore equilibrium disturbed by the deposition of heavier clastics upon a soft plastic sediment or upon the sediment which was in a metastable equilibrium and prone to liquefaction. Suggested processes include: 1) complicated variations in freezing and thawing introduced by differences in grain size and moisture content within a soil body; 2) squeezing of unfrozen soil between a downward freezing active layer and the permafrost table as the active layer becomes anchored to it in some places sooner than others; and 3) density differences during thawing.

Load folds comparable to those described as involutions were produced experimentally by Kindle (1917); Kuenen (1958); and Dzulynskf and Walton (1963). Periglacial

environments provide several favorable conditions which allow these structures to form: 1) abundance of meltwater which maintains saturation at and close to the surface; 2) presence of frozen ground which inhibits the downward escape of water and provides a surface for sliding; 3) scarcity of vegetation; and 4) commonly encountered segregation of soils into layers differing in density and the superposition of loose coarse rock rubble upon water-saturated and clay rich sediment. Subsequent refreezing of the clay-rich deposit may give rise to many forms of deformation.

The contortions observed in the study area are formed as the result of a metastable heavy clastic unit overlying a soft plastic sediment. The process responsible for the formation of the contortions is unknown. Possible processes include variations in freezing and thawing, squeezing of unfrozen soil and density differences between the materials during thawing.

6.2. Pillars

Pillars are defined as a columnar structure perpendicular or oblique to boundary that cuts across laminae or other primary structures in a sand bed, formed by the escape of water after deposition of the bed (Lowe and LoRicols, 1974). The structures typically cross-cut bedding, exhibit vertically-oriented clasts, and have a coarser matrix than the surrounding sediments.

6.2.1. Descriptions

Ravenscrag Site

The Ravenscrag outcrop is located near the southern edge of the Cypress Hills plateau (Figure 5.1) approximately ten kilometers south of the maximum southerly limit of the continental ice sheet. The outcrop consists of planar bedded, well-imbricate gravels (lithofacies 3 and 5) with

one large vertical disturbance to otherwise laterally continuous bedding (Plate 5.9). The pillar structure in the planar bedded conglomerate in Fig. is the largest observed. The pillar is at least 310 cm long and has a maximum width of 90 cm. The interior of the structure consists of a conglomerate with a coarser matrix than the surrounding sediments. Primary bedding within the center of the structure has been modified. Pebbles have a maximum B-axis of 7 cm and no change in maximum grain size (clasts or matrix) was observed within the structure. Many pebbles have their long axes oriented parallel to the pillar axis.

The base of the pillar structure is marked by the interfingering of the sands (Plate 5.9). The sands (lithofacies 9) exhibit sigmoidal shaped climbing ripples, 0.2 to 0.3 cm thick, with elevation differences between the crest and trough of the individual ripples subdued (Plate 5.10). These climbing ripples are only found in the portion of the sand unit which interfingers with the gravels. Elsewhere these sands range from massive to planar cross-bedded. The sands are displaced by small high angle faults and exhibit small water-escape structures (Plate 5.10).

Rockglen Site

The Rockglen section is located two kilometers east of the town of Rockglen in gravel deposits which were not mapped as Wood Mountain Formation by Vonhof (1969) but are lithologically similar. The continental ice sheet reached within five kilometers to the north. The Plateau is bounded to the south by the Rockglen Channel, a Late Wisconsinan meltwater channel (Parizek, 1964). The outcrop is composed of up to 3 m of planar bedded, imbricate gravels of lithofacies 5. The observed pillar structure is 185 cm long and exhibits a maximum width of 40 cm (Plate 5.11). The structure is vertical with fanning at its base and at the top. Within the pillar structure the clasts are generally

Plate 5.9

Pillar exposed at the Ravenscrag Site (inset shows location of Plate 5.10)

Plate 5.10

Sand facies exposed at the base of the pillar at the Ravenscrag Site

Plate 5.11

Pillar exposed at the Rockglen Site

larger and lack the bimodal sorting of the surrounding beds. Vertically oriented clasts are observed. The pillar structure has coarser, and a higher percentage of matrix to clasts as compared to the surrounding beds. The base of the pillar structure forms a layer with similar characteristics to that of the pillar. This basal unit has a coarse grained matrix with unimodal sorted, large clasts and is in some places open worked.

6.2.2. Interpretation

Excluding water escape, Postma (1983) describes three processes which would also produce similar structures: 1) scour and fill; 2) loading; and/or 3) bioturbation. Other processes include faulting and thermal contraction cracking. Scour and fill structures are characterized by their three-dimensional shape and internal cross-bedding. Application of the criteria of Postma (1983) eliminates scour and fill structures as a possible origin for the pillars described here. This is based on these characteristics as well as the variety of sedimentary structures present and the vertical orientation of pebbles in the fine-grained matrix. Scour and fill can be eliminated as a possible origin for the pillar structures observed in this study based on these criteria.

Formation of these structures by loading of fine-grained sediments (Lowe, 1975) will also affect the sediment near the structure. The undisturbed nature of the sedimentary structures near these features makes formation by loading unlikely.

Some bioturbation activity may resemble pocket and pillar like structures although in general the variability of the pockets and pillars compared with the more regular shaped

burrows will usually provide sufficient criteria to distinguish between the two types of structures (Postma, 1983).

Similarity to faults

The downturning of strata adjacent to the pillar structure resembles drag along a fault (fault drag folds or distributive faulting). Fault-related drag, however, has opposite sense on either side of the fault. In all cases the deformation associated with the pipes described has been the same sense (both upturning or both downturning) on both sides of the pipe. This, coupled with the fact that there is no net displacement of strata across the pipe, eliminates the possibility that they originated entirely through faulting.

Thermal contraction cracking

Thermal contraction usually occurs in areas of continuous permafrost but seasonal frost cracking may also occur in non-permafrost areas due to severe winter freezing. Ice wedges, the result of thermal contraction cracking, are rarely more than 4 m deep (Black, 1976; French, 1976), and the depths of seasonal frost cracks are less than 1 m (Washburn, 1973). The pillar structures observed are less than the maximum size for ice wedges (3.1 m), but in general, the older the ice wedge, the deeper it is (Black, 1976). The pillars are relatively narrow and they are isolated, occurring as single structures. If the pillars were formed by thermal contraction cracking, they would represent only the early stages of wedge growth.

Ice wedges usually occur in ice-rich fine-grained (silts and clays) sediments, although they can also be found in coarser sediments but are usually not as well developed. The general morphology for ice wedges is a vertically

oriented triangle, wide at the top and closing to a point at its base. The infilling sediment in wedges is either wind-blown silt and sand (sand wedges) or washed in sand with a lower percentage of gravel, stratified with ice (ice wedges) (Pewe et al, 1969). After melting of the ice, the infilling material may be vertically stratified and contain vertical clasts. Deformation of the surrounding sediments is also a common feature for wedges. The horizontal expansion of inter-wedge perennially frozen ground as it warms during the summer results in the upturning of frozen sediments near ice wedges. As the ice melts, the upbending of the bedding near the wedge plus the disturbance created by the collapse of adjacent sediments may form contorted structure near the wedge border (Pewe et al, 1969).

Although the structures observed at the Ravenscrag and Rockglen sections exhibit upturning and downturning of adjacent strata, and vertical alignment of clasts, they are not believed to be ice wedge casts. The ice wedge hypothesis would not adequately explain the high-angle climbing ripples at the base of the Ravenscrag pillar, nor the fan-shaped morphology of the bases of both pillars. Instead of displaying a V-shaped morphology (as expected with ice wedges), the pillars become wider at the base.

The observed high angle climbing ripples have only been documented under conditions of substantial fallout of sediment from suspension (Harms et al, 1982). Postma (1983) conducted a simple experiment with water running through a tube. The water, with a sudden increase in pore-water pressure, causes local remolding of the sediment and the formation of a pocket or pillar structure.

Water Escape Origin for Pillars

A breakdown of the fabric of the material to such a degree that the grains are no longer mutually supported but

become temporarily separated and dispersed in a pore fluid is termed liquefaction (Allen, 1984). Fluidization is defined as a mechanism that demands that fluid be forced upward through the grain-mass until the downward acting immersed particle weight is balanced by fluid drag (Allen,1984). Externally supplied fluid is essential to fluidization, whereas if a fluid enters a grain-mass during liquefaction, it does so passively. Fluidization, in the strict sense, can never occur in material with a wide grain size variation and irregular shapes because before the largest particles can be supported, the smallest particles are lost by elutriation (Wilson, 1980).

The formation of water escape structures depends on the external influences acting on the sediment, and the internal sediment properties (Lowe, 1975). External influences such as earthquakes, cyclic oscillations (waves, tides, etc.) and rapid sediment accumulation may bring the sediment into a liquefied state (Moore, 1961; Dill,1967; Andresen and Bjerrum, 1967). Sediment properties such as grain size, sorting, composition (presence of phyllosilicates) and permeability largely determine whether or not structures are produced and what kind they are.

Rapidly deposited sediments and coarse grained mass flow deposits generally have a high porosity and loose packing. Expulsion of pore-water by compaction is greatest from these poorly consolidated sediments and the chance of liquefaction and fluidization is great. Renewed rapid sediment accumulation may cause a sudden increase of the pore-water pressure in the underlying beds. The pore-water pressure will be even higher if less permeable mud layers or ill-sorted immature sandy layers are present. A sufficiently high pore-water pressure will locally reduce the frictional resistance between the grains which promotes the escape of pore-water.

Liquefaction of sediments can occur at various times: 1) during syndepositional or early post-depositional stages; 2) during later post-depositional stages as a consequence of loading during deposition of the succeeding bed; or 3) at much later post-depositional times due to sudden loading or as a consequence of overburden weight and compaction of sediments (Hein, 1982). Most water-escape structures form concurrently with or immediately following deposition (Lowe, 1975). These structures are interpreted as post-depositional, and are relatively uncommon.

The conditions necessary to allow the build up of pore-fluids sufficiently to liquefy and fluidize gravels could be met on 10 to 20° depositional slopes (Postma, 1983). Rapid deposition and aggradation and high flow rates may allow these conditions to form on gentle slopes (Johnson, 1986). The Cypress Hills Formation has been interpreted as having a low, >1° depositional slope (Vonhof, 1969), but high flow rates and rapid deposition are inferred from the interpreted episodic flash floods capable of transporting boulder-sized gravel (Leckie and Cheel, 1989). It is possible to create water escape structures syndepositionally based on the interpreted depositional environment of the Cypress Hills Formation.

The water escape structures observed are not believed to have formed syndepositionally due to their distribution. The structures are only observed in those portions of the Cypress Hills Formation that are relatively close to the Late Wisconsinan ice maximal position. These structures are only observed in the upper few meters of the Cypress Hills Formation and are not observed in the cemented portions.

Lowe (1975) states that the upward flow of sediment and water within fluidized channels is commonly accompanied by more or less continuous subsidence of surrounding grain-

supported sands and occasionally slumping of the pillar walls. This slumping produces downturning of the sediments immediately adjacent to the pillar walls. Plint (1983) described downturning in fluidization pipes in Eocene sands. He noted that the downturning occurred as a series of microfaults. Gill and Kuenen (1958) illustrated downturning of beds at the margins of sand volcanos. It is likely that the downturning of beds is the result of upward movement of escaping pore-water. This flow would have been relatively dilute, and failed to support the adjacent walls. This would account for the downturning of beds at the Rockglen pillar. The Ravenscrag pillar exhibits upturning of strata adjacent to the pillar. This could be the result of the flow not being able to remove the larger clasts from the central portion of the pillar. These larger clasts would become reoriented and may be able to provide a framework preventing collapse of adjacent sediments into the pillar.

The pillar structures investigated here are therefore interpreted as being similar in origin to the hydrodynamic blowout structures observed by Christiansen (1982). Hydrodynamic blowout structures are the result of the upward discharge of groundwater from an underlying confined aquifer through naturally or artificially produced discontinuity in the overlying confining beds (Christiansen, 1982). The inferred conditions necessary to produce hydrodynamic blowout structures are: 1) the area is underlain by an extensive artesian aquifer capable of delivering at least $4m^3/s$; 2) the hydrostatic pressure in the aquifer approaches or exceeds the critical uplift or lithostatic pressure; and 3) a local discontinuity must be present to initiate venting through the overlying beds.

Although the Cypress Hills Formation is not an extensive aquifer at present, conditions during the Late Wisconsinan may have produced similar results. At this time, permafrost

conditions were present (Jungerius 1966, Westgate 1965). The frost penetration through a gravel layer would be deeper than through a finer grained sediment. Thus the base of the active layer is interpreted as being near the base of the gravel or, where the unit is cemented, at the top of the cemented portion. Frozen ground is not an impermeable barrier but instead a confining material with very low hydraulic conductivity (Sloan and van Everdingen 1988). At this time, it is highly probable that the gravels were water saturated due to the proximity of the ice margin, the wash effect on these surfaces (R.W.Klassen,Geological Survey of Canada, personal communication, 1990), and the lacustrine sediments on adjacent surfaces (Kulig, in prep.). This would allow for the first condition outlined by Christiansen (1982) to be met.

The second condition, as described by Christiansen (1982), is met by similar circumstances as prevailed at Howe Lake. Since water saturated horizons were present within the sedimentary sequence that extended beneath the ice complex, the pore-fluid pressure in these water-saturated layers would be increased by an amount approaching that produced by the additional load of the complex. With a recharge upland set up by the position of the ice margin, there would be a strong tendency for flow away from the ice position and upward discharge to occur near the stagnant ice margin (Christiansen, 1982). The actual vertical escape can form in two ways. The first mechanism would involve conditions similar to those in which a pingo would form. During the cold season of the year, freezing extends downward toward the permafrost table. This confines any water present if there is no permeable unfrozen zone available for relief of hydrostatic pressure downward or laterally. In pingos, the water may be subjected to such great hydrostatic pressure that it ruptures the seasonally frozen layer (Williams, 1970).

The pressure of confined water freezing is sufficient to cause explosive rock-bursts, a brittle fracture induced by the sudden release of stored strain energy (Taylor, 1963). Rockbursts occur in rocks that can store enough strain energy to fail in a brittle fashion. This can occur as a saturated block undergoes freezing from the outside. As the water freezes, rock pores become plugged by ice from the surface. Davidson and Nye (1985) have shown that an ice seal of rock pores can sustain pressures as high as 10MPa. Although dissipation through seepage is greater in unconsolidated sediments than in rocks, this is an indication of the strength of ice.

Clayton and Moran (1974) have outlined two groundwater pressure zones, 1) a zone of persistent excess pore-water pressure, and 2) a zone of potential dissipation of excess pore-water pressure. The zone of persistent excess pore-water pressure roughly approximates with the boundaries of the thawed bed zone. Under conditions of fine-grained sediments overlying a buried aquifer, and poor drainage (similar to the deforming glacier bed model as outlined by Boulton and Jones (1979)), coupled with high loading from overlying ice, the result will be elevated pore-water pressure (fine-grained sediment is more compressible than coarse grained sediment), and decreased basal shear strength of glacier bed sediments. The zone of potential dissipation of excess pore-water pressure roughly corresponds to the frozen bed zone. This boundary may move up ice (into the thawed bed zone) as subglacial conditions change, however.

The possibility of the water escaping horizontally may be a problem. At present, the cliffs to the south are cemented. Kupsch and Vonhof (1967) interpret the cement as a post-depositional and post-erosional feature. They believe that the occurrence of well-cemented gravels in valley cut in the Pleistocene indicates that the cementation

process is post-glacial. The lateral extent of the cementation is small and the process is continuous with slope retreat (Kupsch and Vonhof, 1967). This would indicate that cementation may have been present prior to glaciation, on valley walls which were present at this time. The sediment supplied by the glacier, and thus available in the groundwater for cementation, might contain calcareous material but most carbonates are derived from solution of calcite and dolomite occurring in the Tertiary sequence (Kupsch and Vonhof, 1967).

The suggestion that extreme confinement to create high pore-water pressures as required may be unfounded. Burbidge et al (1988) interpreted fissures in glaciofluvial sediments as being formed by water-escape processes. The sediments were charged with high pore-water pressures by remnants of glacial ice buried beneath the sediments exhibiting water escape structures. They believe that upward relief of pressure might initially take place by seepage, followed by concentration of seepage into planar channels. The velocity of seepage flow eventually surpassed that necessary for fluidization, and the formation of water escape structures was initiated.

The pillar structures observed in the study area are interpreted as being the result of the upward discharge of groundwater from an underlying aquifer through a discontinuity in the overlying confining beds. The confinement is provided by the seasonal freezing of the ground.

6.3. Reoriented Units

Although the majority of these units were identified as reoriented in the field, the clarification of this classification was based upon the compilation of the fabric data. The criterion used to classify deposits as reoriented

is the average dip of the clasts must be greater than 60° (Plate 5.12). The use of 60° was chosen arbitrarily as fabric data from previous literature indicated that primary fabric orientation exhibit dips up to 45° (average). Due to possible natural scatter or error (bias) in sampling, a 15° margin of error was allowed creating an artificial division between "normal" gravels and "reoriented" gravels. This division is supported by the distinct lack of clasts with dips in the 45° to 60° range.

6.3.1.Descriptions

The description of the reoriented gravels falls within two distinct lithofacies, lithofacies 2 and lithofacies 11. Lithofacies 2, structureless matrix supported gravel encompasses all of the reoriented locations which were below the present surface as well as some occurring at the surface. Lithofacies 11 always occurs as the uppermost facies within the observed sections, when present. With the exception of lithofacies 11, massive silts with rare stones, all reoriented clasts occurred in lithofacies 2, but not all occurrences of lithofacies 2 contained reoriented clasts.

6.3.1.1 Matrix

The parts of lithofacies 2 classified as reoriented contain 40-70% matrix. This matrix is a mix of sand, silt and clay, with concentrations of 30-80% sand, 16-52% silt, and 2-41% clay. There appears to be a correlation between the amount of silt and the average clast dip (Figure 5.2). The closer the average clast dip to 90°, the higher the proportion of silt in the matrix. There is also a rough correlation between the amount of fines (silt and clay) in the matrix and the average dip of the clasts (Figure 5.3). Their relationship is linear, as with the silt alone, but there appears to be a limiting amount of clay. Once the

clay reaches a certain proportion, the average clast dip continues to increase but at a much slower rate.

This relationship between matrix texture is also illustrated in a ternary diagram (Figure 5.4). This ternary diagram illustrates the percentage of sand, silt and clay in the matrix for all occurrences of lithofacies 2. Lithofacies 2 is divided into reoriented and normal samples. Those samples with the average dip of the clasts greater than 60° contained a higher proportion of silt and clay compared to those samples with the average dip of clasts less than 45°. The relationship exhibited on the ternary diagram shows a transition zone which may be explained by factors other than matrix texture.

Clay mineralogy of the matrix was not examined. Taylor (1987) found that clay separate analyses conducted on the Cypress Hills Formation were consistent with a theory of uniform smectite neogenesis throughout a vertical profile within the Horseshoe Canyon section (Figure 5.1). The elemental, surface area, cation exchange capacity and x-ray diffraction analyses all showed a general uniformity in clay mineralogy with depth.

Plate 5.12

Typical exposure of the reoriented units

Figure 5.2

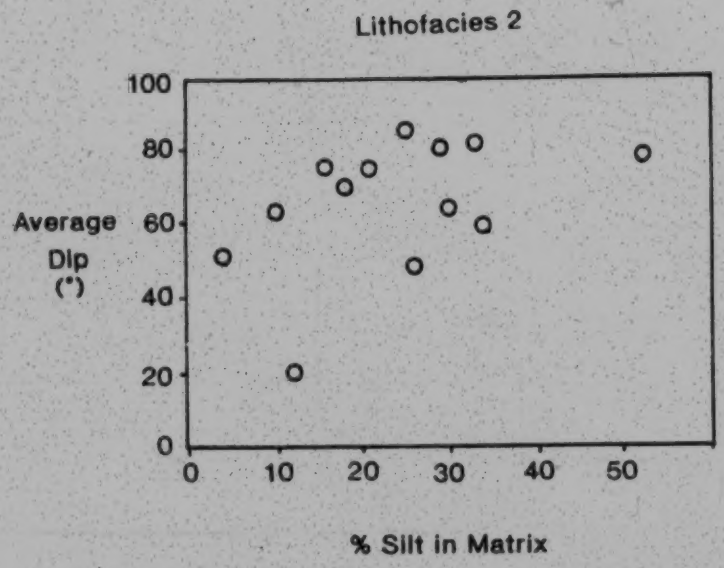

Lithofacies 2
Average Dip (°)
100
80
60
40
20
0
0
10
20
30
40
50
% Silt in Matrix

Figure 5.3

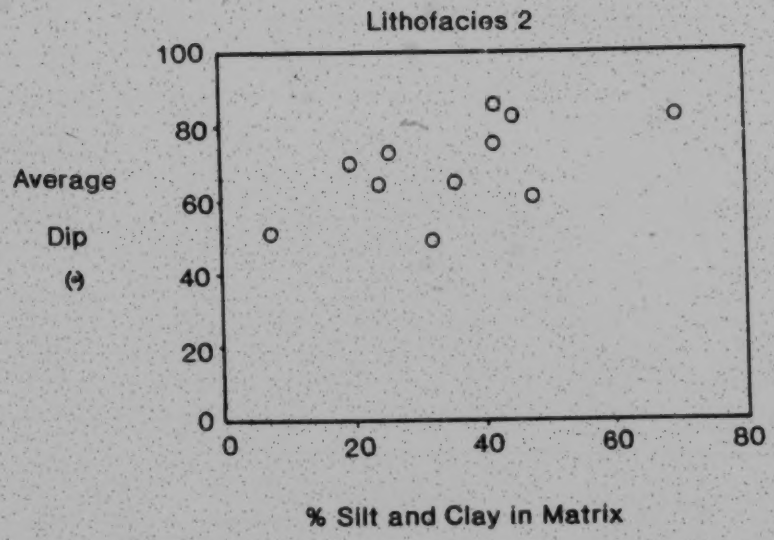
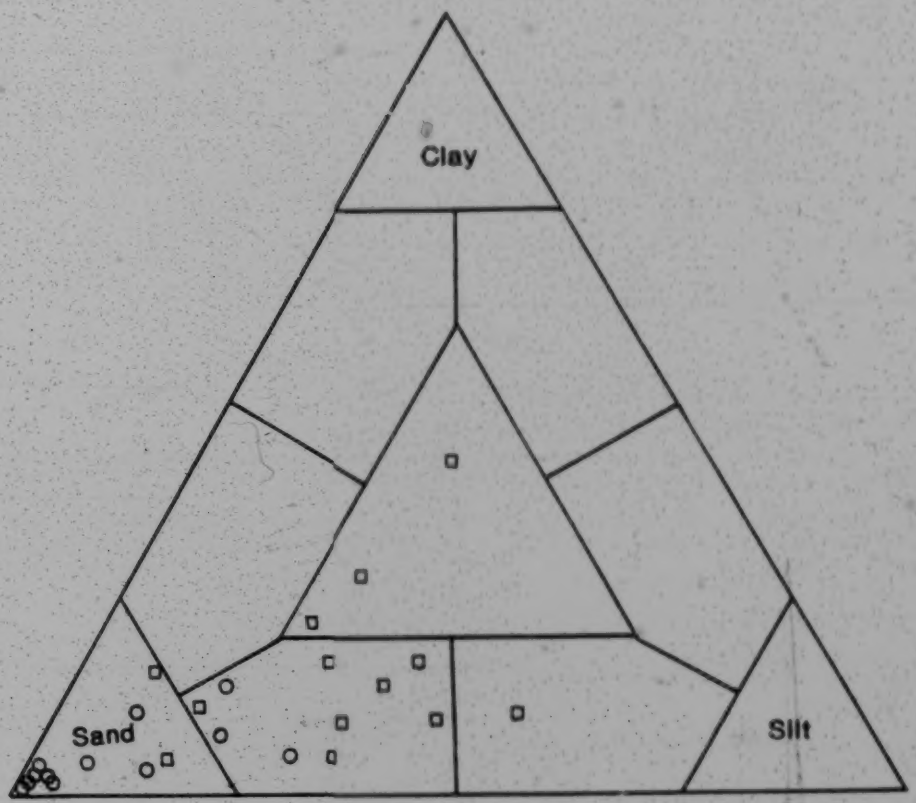

Lithofacies 2
Average
Dip
(°)
100
80
60
40
20
0
0
20
40
60
80
% Silt and Clay in Matrix

Figure 5.4

Textural Comparison of Similar Depth Samples with Varying Average Clast Dip



○ Average clast dip <45

□ Average clast dip >60

6.3.1.2 Clasts

The clasts are dominantly discose to roller, well-rounded with lesser amounts of blade and equant shapes present. The clasts are moderately sorted comprised largely of pebbles and cobbles with some small boulders present. There was a significant difference observed in the shapes of those clasts oriented 60° to 90° compared to those clasts dipping less than 45° (Figure 5.5). Little can be concluded regarding the shape of the clasts dipping between 45° and 60° as there were not enough observed to comprise a statistically significant population. The population of clasts dipping less than 45° are of all shapes, with the majority being classified as rollers. The mean A/B axes ratio is 0.600 and the mean B/C axes ratio is 0.698 (Figure 5.5). This is noticeably different from the population of clasts dipping 60° to 90°. This population has a mean A/B axes ratio of 0.683 and a mean B/C axes ratio of 0.512 (Figure 5.5) classifying the mean clast shape as discose.

6.3.2.Interpretation

A dominantly vertical plunge of pebbles is not a primary orientation. Imbricated fluvial deposits commonly show average dips of 10° to 30°. Primary beach deposits have exhibited dips up to 45° (Krumbein, 1939). Some beach deposits have been described with vertical clast orientations.

Irregular clast distributions with many clasts exhibiting vertically oriented a-axes have been observed in many debris flow deposits (Bull, 1970; Pierson, 1980). These are usually viscous flows with lower water contents. Kulig (1985) conducted fabric analysis on ten deposits he interpreted as debris flows. Despite observing many

vertically-oriented clasts, the average plunge of the clasts within these deposits ranged from 0.3 to 25°. The debris flow deposits exhibited a multimodal clast distribution.

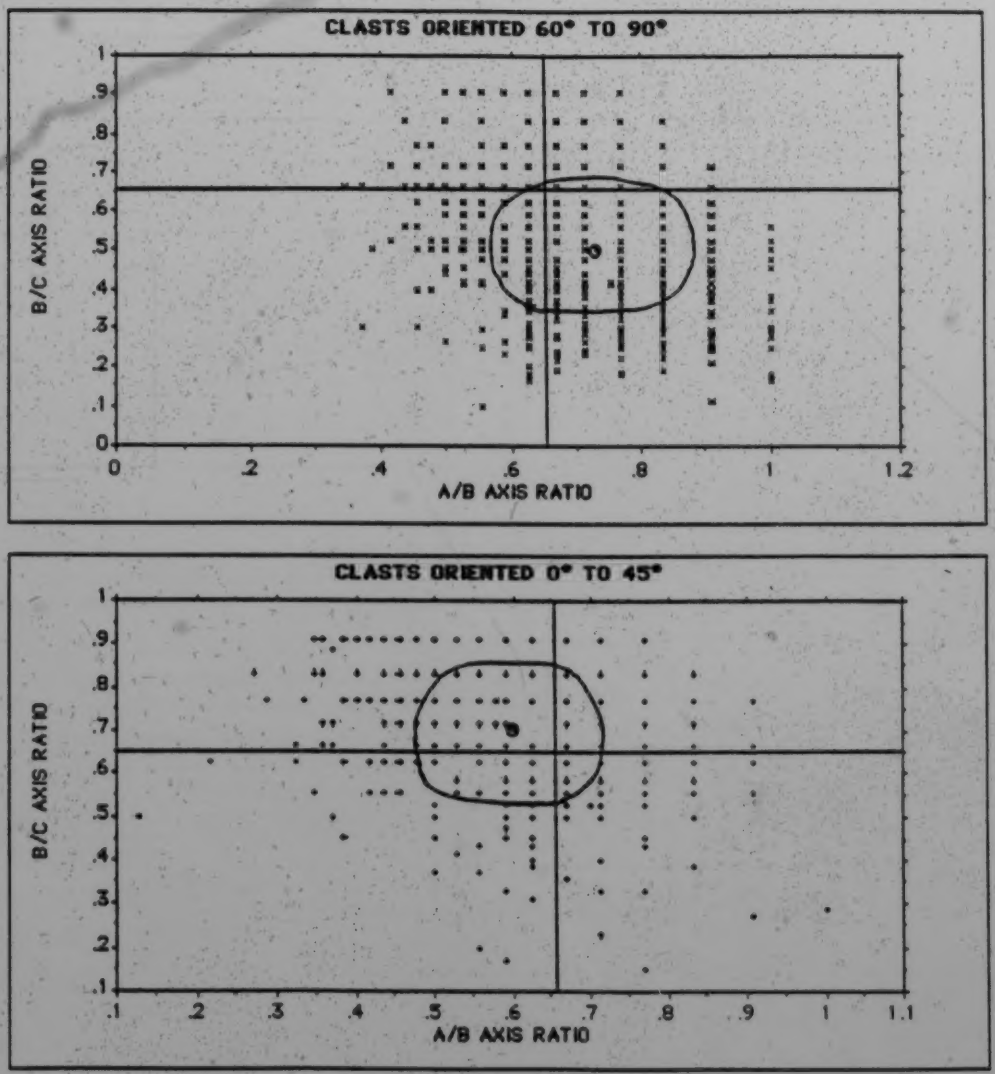
The reoriented units have characteristics similar to debris flow deposits with the exception of the fabric orientations. There reoriented portions have primary descriptions corresponding to those of lithofacies 2, structureless matrix supported gravel. This lithofacies has been interpreted as rapid deposition from turbulent flows. The poorly developed fabric may be the result of insufficient time to create an organized fabric, among other factors. The interpretation of these deposits does not adequately explain the high plunge of the clasts present. The "normal" deposits of lithofacies 2 exhibit clast dips typical for debris flow deposits but the "reoriented" portions cannot be the result of primary deposition by debris flow processes without any subsequent modification.

The presence of pebbles with long axes oriented vertically has often been considered evidence for permafrost conditions. Westgate (1968:31) interpreted the upper portion of the Cypress Hills Formation as "deformed by frost action, as is evidenced by the presence of deranged pebbles". Feltham (1989) interpreted vertically oriented pebbles as the result of periglacial frost action.

Whether the rotation occurs during freezing or whether it occurs during thaw is unknown and not important here, the important aspects are that in order for rotation to occur, segregated ice lenses must be able to form.

The two most important factors in determining whether segregated ice lenses will form in a sediment are its texture and pore size. Brandl (1970) found that the amount and size distribution of the matrix, given that the gravel is matrix supported, are critical factors. It is generally

FIGURE 5.5


CLASTS ORIENTED 60° TO 90°
B/C AXIS RATIO
A/B AXIS RATIO
0
.2
.4
.6
.8
1
1.2
CLASTS ORIENTED 0° TO 45°
B/C AXIS RATIO
A/B AXIS RATIO
.1
.2
.3
.4
.5
.6
.7
.8
.9
1
1.1
mean
standard deviation

agreed that if the deposit contains more than 8% of material less than 0.02 mm in diameter the deposit must be considered as susceptible to segregated ice lens growth.

The amount of matrix within the reoriented units was determined to be a critical factor. All of the reoriented units examined had a minimum of forty percent matrix and all could be described as matrix-supported. This is important in that within all the outcrops exposed within the study area, the matrix-supported gravels could be considered infrequent to rare, comprising approximately five percent of all gravels examined. This minimum amount of matrix required for the reorientation process to occur indicates that the clasts must be "free" to rotate. Clast to clast contacts would inhibit the reorientation process.

The texture of the matrix was also found to be an important factor. There is correlation between the percentage of silt in the matrix and the average dip of the clasts in the sample. The higher the percentage of silt in the matrix, the more vertical the average clast dip, to a maximum percentage of silt of 40% to 50% (Figure 5.2). This may be an apparent maximum as the average clast dip reached over 80° and due to natural scatter, this may be the absolute maximum expected average.

The importance of silt in the matrix may be evidence for the reoriented units due to frost action. Silt is particularly prone to frost heaving as it permits relatively easy migration during freezing and thus will form segregated ice lenses. Clay has little permeability and as such will not develop ice lenses to the same degree as silt will. Coarse sands and gravels have a greater permeability than silt and permit higher rates of migration of water, but lack the suction potential of silt. The potential for drawing water to the freezing front increases with decreasing grain

size. Silt is the most susceptible grain size for segregated ice lenses to form.

However the presence of silt is not convincing evidence for a frost action origin as it may also be used as evidence for a liquefaction/fluidization origin. Lowe (1975) outlines the most important factors known to influence liquefaction in sediments, one being grain size. He suggests that fine-grained sands and coarse-grained silts tend to show a greater susceptibility to liquefaction than coarse-grained sand and gravel, or finer-grained silt and clay. In general, both coarse sand and gravel, and fine silt and clay dewater by seepage. The importance of silt may be enhanced by the presence of sand in the matrix. The reoriented units contained between 30% and 80% sand in the matrix. Lowe (1975) states that a strongly bimodal sediment, such as a gravel, is likely to show fluidization of the smaller particle population within the pores of the undisturbed larger grains. In this way, clay or silt may be fluidized and elutriated within the pores of a coarser sand or gravel while the overall grain framework remains intact. If we consider that with a larger amount of matrix that the matrix may become fluidized, the larger grains may be disturbed, rotating to a vertical position to offer less resistance to the escaping water.

The amount of silt was not the only factor observed to influence the amount of reorientation observed in the samples. The percentage of silt and clay in the matrix also influences the average clast dip (Figure 5.3). As with the percentage of silt previously discussed, the more silt and clay within the matrix, the more vertical the average clast dip. There appears to be a limiting amount of clay that can be present in the matrix. As the clay reaches 20% of the matrix, the average clast dip increases at much slower rate than when the amount of clay is less than 20%. This may also

be due to a relative decrease in the percentage of silt as the percentage of clay increased.

The discussion regarding possible interpretations with reference to the amount of silt and clay are the same as discussed for the percentage of silt in the matrix.

The pore size is related to grain size and the grain size distribution of the soil. Silt permits relatively easy migration of moisture during freezing and thus favors ice lens growth.

The theory of clast reorientation due to segregated ice lens growth is not without problems. Burrous (1978) noted that all originally inclined objects emerged from the soil at 45° and that all rotation occurred after the objects had emerged from the soil. This is similar to Pissart's (1973) experiments as the segregated ice lenses formed at the surface. Washburn (1980) illustrated the reorientation process in a related way. The rotation is in response to frost heaving (and thus the formation of segregated ice lenses) at the ground surface.

Burrous (1978) conducted experiments on the influence of freeze-thaw cycles on different clast shapes. The three shapes he used, discs, laths, and spheres, did not produce any significant differences in either heave rate or orientation change. Although this may not be conclusive due to factors such a chosen shapes, clast composition, matrix amount or composition, it is remarkably different than the orientations with reference to shapes noted within the reoriented units. The dominant clast shape within the 'normal' debris flow unit is classified as a roller. Rollers comprise 37% of the clasts which were measured as having dips of 45° or less and 8% of the clasts measured as having dips of 60° or greater. Disc shaped clasts comprised 20% of the clasts oriented 0° to 45° and 73% of the clasts oriented

60° to 90°. 15% of the clasts with dips of 60° to 90° were classified as blade shaped.

The high percentage of clasts oriented 60° to 90° were measured as having a b/c axes ratio of less than 0.66. This would indicate that the b-axis is significantly larger than the c-axis. This shape difference may indicate that pressure is responsible in some way for the reorientation process. The larger b-axis would have displayed a larger surface which may have inhibited the sediment from flowing around the clast. The flow of sediment around the clast is believed to be a major factor in the upheaving of clasts due to frost action (Prestrud, 1987). This larger surface area may have prevented the flow of sediment around the clast and instead caused a reorientation into such an orientation that the clast offers least resistance to mobile sediment. Given that the pressure gradients were vertical in the sediment, the clast's orientation which would offer the most 'aerodynamically' stable position would be vertical. These pressure gradients need not invoke permafrost to form. Soft sediment deformations form as a result of such pressure gradients.

Many of the units discussed here have non-planar bases. These contacts are described within the section dealing with contortions. The irregular nature of the contacts forms structures which are termed pocket structures. Postma (1983) interpreted pocket structures as forming from the liquefaction of sediment induced by sudden increases of pore pressure. During liquefaction, temporary support of sediments by escaping pore fluids creates decreased frictional resistance between grains so that pebbles sink downward because of their larger size and finer grains are displaced upward. Pebbles sink downward to the bottom of the liquefied zone forming a pocket structure. While sinking, pebble long axes may rotate into a vertical

orientation in order to offer the least resistance to upward escaping waters. The intimate association between the reoriented units and the irregular basal contacts would support the possibility of water-saturated sediments with high pore pressures causing or aiding the reorientation of the clasts.

The reoriented units observed in the study area are interpreted as being the result of liquefaction of the sediment. Rotation of the clasts by the formation of ice lenses in a permafrost environment is also a possible interpretation but the previous interpretation of the contortions and the pillars does not support this. As contortions form as the result of an instability and a similar process forms the pillar structures, it follows that a similar process is most likely responsible for the reoriented units.

Chapter 7
Distribution of Secondary Sedimentary Structures

Three types of secondary sedimentary structures, contortions, pillars, and reoriented units have been described and interpreted in Chapter 5. The location of the sections which contain these structures and the glacial limits are shown in Figure 1.2. The observed locations of these structures will be discussed in relation to the mapped ice limits and the distribution of the primary lithofacies to aid in the interpretation of the processes responsible for the formation of the secondary structures observed.

7.1. Distribution of Contorted Units

The units which displayed contortions were observed at four sections within the study area (Figure 1.2). These are all located beyond the Late Wisconsin ice limit. The contorted units appear to be widespread, occurring near both the western and eastern limits of the study area.

The controls in determining the distribution of the contorted units are the ice limits and the primary lithofacies distribution.

The Late Wisconsin ice limit influences the distribution of the contortions as all the observed contorted units occur outside the area effected by Late Wisconsin ice. Other contortions may have developed within the limits of the Late Wisconsin ice but were subsequently eroded as the ice advanced.

The primary lithofacies distribution is important in controlling the distribution of contortions. A clast supported unit was observed associated with a matrix supported unit. This association is necessary to form the

density differences and/or variations in thermal conductivity which are required to form the contortions.

The Cypress Hills Formation has been interpreted as a semi-arid alluvial plain (Leckie and Cheel, 1989). This depositional environment is characterized by a rapid decrease in grain size and an increase in proportion of finer facies distally. It would be expected that the distribution of contorted units be more prominent in proximal areas, and thus in the western portion of the study area. It is difficult to confirm this as only four contorted sections were observed, but three of the four sections are located in the western portion of the study area.

7.2. Distribution of Pillars

Within the study area, two sections displaying pillars were observed (Figure 1.2). The pillars have been interpreted to have formed from rapid escape of water confined within the sediments. The upper confinement is believed to have been provided by the freezing of the seasonal active layer. The base of the pillars is marked by a layer of moderately to well cemented sands. Lateral confinement may also be provided by cementation.

The pillars were found in two sections, Ravenscrag and Rockglen. Both of the sections are located beyond the Late Wisconsin ice limit. The Ravenscrag section may have had ice as close as 3 km to the south in the present Frenchman River valley. The Rockglen section is on a bench bounded to the south by a meltwater channel. Parizek (1964) mapped an open meltwater channel or spillway 2 km south of the Rockglen section. Parizek interpreted this channel as forming during or immediately following the Late Wisconsin ice maximum.

7.3. Distribution of Reoriented Units

Reoriented units are widespread within the study area, being observed in both the westernmost and the easternmost sections examined. Reoriented units were observed at eight locations within the study area. All locations at which the reoriented units were observed were located beyond the limit of the Late Wisconsin ice.

The controls in determining the distribution of the reoriented units are the ice limit and the primary lithofacies distribution.

The Late Wisconsin ice limit is believed to be the control on the distribution as all the reoriented units observed were located beyond this limit. Reoriented clasts may have formed prior to the ice sheet reaching its maximal position if conditions were right. This is not conclusive as it should also be noted that the majority of the gravels belonging to the Cypress Hills and Wood Mountain Formations are located beyond the limits of the area affected by the Late Wisconsin ice.

Reoriented units were defined as a variation of lithofacies 2, structureless matrix-supported gravel. It is believed that all reoriented units were deposited primarily as lithofacies 2 and have undergone subsequent modification. The distribution of lithofacies 2 would thus be a limiting factor in the distribution of the reoriented units. Lithofacies 2 is common in the upper portions of the sections examined and occurs less frequently in the lower portions. Lithofacies 2, although widespread in the study area, was observed more frequently in the western sections than in the eastern sections. This observed distribution of lithofacies 2 would support Leckie and Cheel (1989)'s interpretation of the Cypress Hills Formation as a semi-arid alluvial plain. The source of the gravels for the alluvial plain would be to the west-southwest of the study area.

Debris flows (lithofacies 2) would be more common in the proximal reaches (west) and decrease distally (east).

The distribution of the secondary structures observed is controlled by both the limit of the Late Wisconsin ice as well as the distribution of the primary lithofacies. Another major factor in the distribution of these structures is the former climate. Conditions necessary for any of these structures to form include a climate colder than the present climate in the area. Periglacial conditions were likely present in the study area during and/or just prior to the maximum extent of the ice. These conditions have not existed in the area since the retreat of the ice.

Chapter 8
Conclusions

Stream flow, both in the form of channelized and unchannelized sheetflood, is the dominant, if not nearly exclusive depositional process on humid fans and also accounts for the largest volume overall of deposits on arid-fans. Sediment gravity flows, including debris flows, mudflows, and landslides are important processes on arid fans. The interpreted lithofacies within this study range from deposition as longitudinal or diagonal bars to diffuse gravel sheets to debris flow deposits. The variation in grain size from west to east, a general fining process, coupled with paleocurrent analysis would indicate a westerly source. The primary depositional facies observed can be interpreted as forming in a semi-arid alluvial plain. The most recent interpretation of the Cypress Hills Formation dates its deposition from Late Eocene to Early Oligocene, 44-35 Ma (Leckie and Cheel, 1989), and the Wood Mountain Formation younger. The gravels have changed since their deposition in an alluvial plain to form the caprock of the present plateau.

The Cypress Hills and Wood Mountain Formations exhibit structures which are not the result of normal stream flow processes. These secondary sedimentary structures were formed post-depositionally as the result of other processes acting upon the gravels. These secondary structures can be divided into three types: contortions, pillars and reoriented units.

The contorted units are interpreted as being the result of soft sediment deformations due to conditions set up under permafrost conditions. Processes which occur in permafrost environments which would be favorable for deformations to occur include, density differences set up during thawing of sediment; variations in freezing and thawing induced by

differences in grain size or moisture content of the sediments; and the confining of unfrozen sediment between a downward freezing active layer and the permafrost table. These processes may not occur in isolation but instead the formation of the deformations may be a result of all the processes occurring somewhat simultaneously.

The pillars are interpreted as water escape structures forming in response to high pore water pressures. The high pore water pressures are the result of confinement of groundwater between the permafrost table or a cementation boundary and the downward freezing active layer.

Reoriented units are formed when a water-saturated matrix-supported gravel is confined under high pressures which would lead to the fluidization of the sediment. The confinement is caused by the same factors as the contortions and the pillars. This is the favored interpretation for the reoriented units due to their association with the other secondary structures observed. The Ravenscrag section exhibits contorted units and numerous pillars. Portions of those units which are contorted match the criteria of the reoriented units. All of these structures require permafrost conditions to form.

Conclusions can be reached based on the distribution of the secondary structures and the associated lithofacies:

1) The distribution of debris flow units and therefore the reoriented units changes within the study area. Debris flows occur most frequently in the western sections and in the upper part of the sections. The eastern sections exhibit more sand and well sorted deposits. It follows then that the reoriented units are more prevalent in the western sections than in the eastern sections. As paleoclimatic conditions were probably similar throughout the study area, the scarcity of reoriented units in the eastern sections may be the result of unsuitable primary lithological conditions necessary for their formation.

2) The distribution of primary lithofacies controls the occurrence of the contorted structures. The lithofacies distribution changes from the western sections to the eastern sections from matrix-supported gravels occurring commonly in the west to rare occurrences towards the east. The presence of matrix-supported units leads to density differences during freezing and thawing of the sediment as the clasts are better conductors of heat than the pore spaces within the matrix. The distribution of the matrix-supported gravel units is an important factor necessary for contortion formation was concluded to be a clast-supported gravel in contact with a matrix-supported gravel.

3) The ice limits were found to be useful when considering the distribution of the secondary structures. All three of the structures were found to occur only beyond the Late Wisconsin ice limit. The unglaciated area contained sediments which exhibit contortions and reoriented units. Pillars were only found in the area outside the Late Wisconsin ice limit. Only within this area were the conditions necessary for pillar formation present.

The Center Block section exposes sediments exhibiting reoriented and contorted units which are truncated. The truncated occurs at the base of a large trough of cross-bedded gravels containing clasts derived from the Canadian Shield. The truncation surface is interpreted to have formed when the ice sheet was at its maximum, never topping the Cypress Hills but carving spillways from meltwater. The only places where erratics are found on this bench is within the areas interpreted to be meltwater channels by air photograph analysis and the Center Block section data. The truncation of the secondary structures by the channel indicates that the conditions necessary to form these structures were set up prior to the ice sheet reaching its maximal extent. It is unknown if favorable conditions for secondary structure formation were present after the ice sheet reached it maximum as all the glaciofluvial deposits examined,

including these channels, exhibited clast-supported facies, which are unsuitable for the formation of these structures.

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Appendix 1 Climatic Data

Station Name	Latitude	Longitude (W)	Elevation (feet)	Precipitation (inches)	Temperature (°F)
Del Bonita	49°0'	112°47'	4337	13.40	44.8
Santa Rita	48°59'	112°19'	4090	15.29	45.0
Sunburst	48°54'	111°44'	3610	13.02	47.3
Sweetgrass	49°0'	111°57'	3466	8.99	44.5
Chester	48°31'	110°57'	3165	10.28	41.0
Goldbutte	48°59'	111°24'	3498	12.19	41.7
Rudyard	48°59'	110°35'	2980	12.37	46.8
Simpson	48°59'	110°19'	2740	10.90	40.4
Forks	48°47'	107°28'	2600	10.97	39.8
Morgan	49°0'	107°50'	2830	9.43	46.8
Turner	48°59'	108°23'	2982	9.40	42.6
Whitewater	48°46'	107°38'	2355	8.87	44.7
Opheim I	49°0'	106°23'	2980	10.48	43.6
Opheim II	48°42'	106°19'	2950	12.10	43.5
Plentywood	48°47'	104°33'	2043	13.19	46.2
Raymond	49°0'	104°35'	2350	11.51	46.3
Redstone	48°50'	104°57'	2110	13.30	45.5
Scobey	48°50'	105°26'	2480	11.83	46.5
Westby	48°52'	104°30'	2105	10.42	50.3

(average values for the period 1951 to 1980 inclusive)
Source: Climatological Data, Montana
Department of Commerce,
National Oceanic and Atmospheric Administration
National Environmental Satellite, Data, and Information Service
National Climatic Data Center
Federal Building
Asheville, N.C. 28801

Supplied by: U.S. Customs Service, Cut Bank, Montana

Cypress Hills Provincial Park Climatic Information

Year	Min. Temp. (°C)	Max.Temp. (°C)	Precipitation (mm)
1983	-40	31	545.1
1984	-34	31	395.1
1985	-35	29	522.6
1986	-31	27	718.6
1987	-27	30	430.1
1988	-30	30	327.6

Source: Alberta Forest Service Weather Station Reports,
Elkwater, Alberta

courtesy of: T.N. Trott

Appendix 2

Section Locations

Name	Location	Map Sheet
Horseshoe Canyon	NW9-3-8 W4M	72E
Police Point	N14-1-8 W4M	72E
Center Block	NE19-27-8 W3M	72F
Ravenscrag	SW22-23-7 W3M	72F
Resch	SE4-7-6 W3M	72G
Wideview	NW14-5-6 W3M	72G
St. John's	NE30-2-30 W2M	72G
Rockglen	SW1-3-30 W2M	72G